AL-Mustansiriyah University
College of Science
Atmospheric Sciences Department

Remote Sensing Lectures

For 3rd Class Students

By:

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References

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1.What is Remote Sensing?
Remote sensing is the science (and to some extent, art) of acquiring information about the Earth's surface without actually being in contact with it. This is done by sensing and recording reflected or emitted energy and processing, analyzing, and applying that information." In much of remote sensing, **the process** involves an interaction between incident radiation and the targets of interest. This is exemplified by the use of imaging systems where the following seven elements are involved. Note, however that remote sensing also involves the sensing of emitted energy and the use of non-imaging sensors.

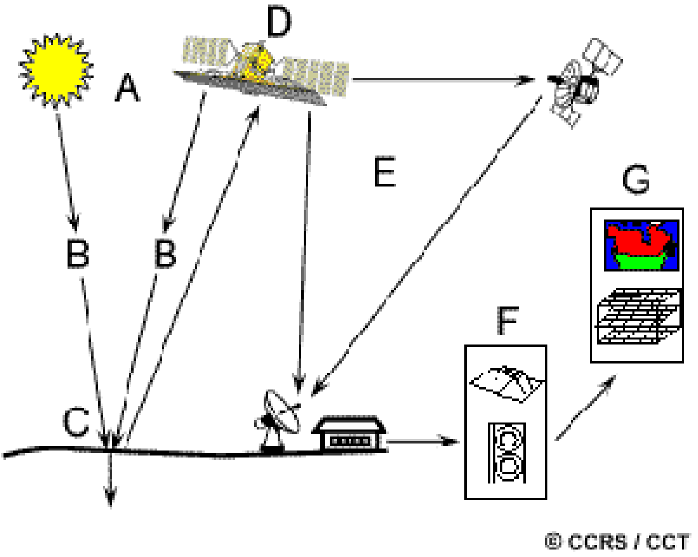


Figure (1): *process of Remote Sensing.*

1. Energy Source or Illumination (A) – the first requirement for remote sensing is to have an energy source which illuminates or provides electromagnetic energy to the target of interest.
2. Radiation and the Atmosphere (B) – as the energy travels from its source to the target, it will come in contact with and interact with the atmosphere it passes through. This interaction may take place a second time as the energy travels from the target to the sensor.
3. Interaction with the Target (C) - once the energy makes its way to the target through the atmosphere, it interacts with the target depending on the properties of both the target and the radiation.
4. Recording of Energy by the Sensor (D) - after the energy has been scattered by, or emitted from the target, we require a sensor (remote - not in contact with the target) to collect and record the electromagnetic radiation.
5. Transmission, Reception, and Processing (E) - the energy recorded by the sensor has to be transmitted, often in electronic form, to a receiving and processing station where the data are processed into an image (hardcopy and/or digital).
6. Interpretation and Analysis (F) - the processed image is interpreted, visually and/or digitally or electronically, to extract information about the target which was illuminated.

7. Application (G) - the final element of the remote sensing process is achieved when we apply the information we have been able to extract from the imagery about the target in order to better understand it, reveal some new information, or assist in solving a particular problem.
These seven elements comprise the remote sensing process from beginning to end.

1.2 Passive vs. Active Sensing

So far, throughout this chapter, we have made various references to the sun as a source of energy or radiation. The sun provides a very convenient source of energy for remote sensing. The sun's energy is either **reflected**, as it is for visible wavelengths, or absorbed and then **reemitted**, as it is for thermal infrared wavelengths. Remote sensing systems which measure energy that is naturally available are called **passive sensors**. Passive sensors can only be used to detect energy when the naturally occurring energy is available. For all reflected energy, this can only take place during the time when the sun is illuminating the Earth. There is no reflected energy available from the sun at night. Energy that is naturally emitted (such as thermal infrared) can be detected day or night, as long as the amount of energy is large enough to be recorded.

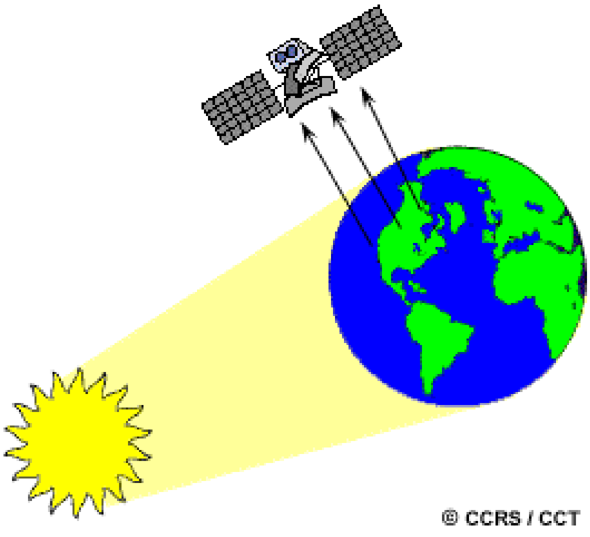


Figure (2): Passive remote sensing

Active sensors, on the other hand, provide their own energy source for illumination. The sensor emits radiation which is directed toward the target to be investigated. The radiation reflected from that target is detected and measured by the sensor. Advantages for active sensors include the ability to obtain measurements anytime, regardless of the time of day or season. Active sensors can be used for examining wavelengths that are not sufficiently provided by the sun, such as microwaves, or to better control the way a target is illuminated. However, active systems require the generation of a fairly large amount of energy to adequately illuminate targets. Some examples of active sensors are a laser fluorosensor and a synthetic

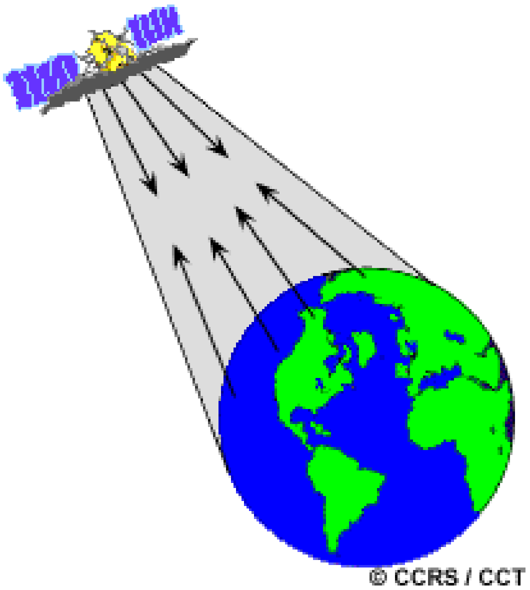


Figure (3): Passive remote sensing

1.3 Sensors

A sensor is a device that measures and records electromagnetic energy. Sensors can be divided into two groups. Passive sensors depend on an external source of energy, usually the Sun (although sometimes the Earth itself). The group of passive sensors cover the electromagnetic spectrum in the range from less than 1 picometre (gamma rays) to over 1 metre (micro and radio waves). The oldest and most common type of passive sensor is the (photographic) camera. Active sensors have their own source of energy. Measurements by active sensors are more controlled because they do not depend upon the (varying) illumination conditions. Active sensors include the laser altimeter (using infrared light) and radar. Figure 3 gives an overview of the types of the sensors that are introduced in this section.

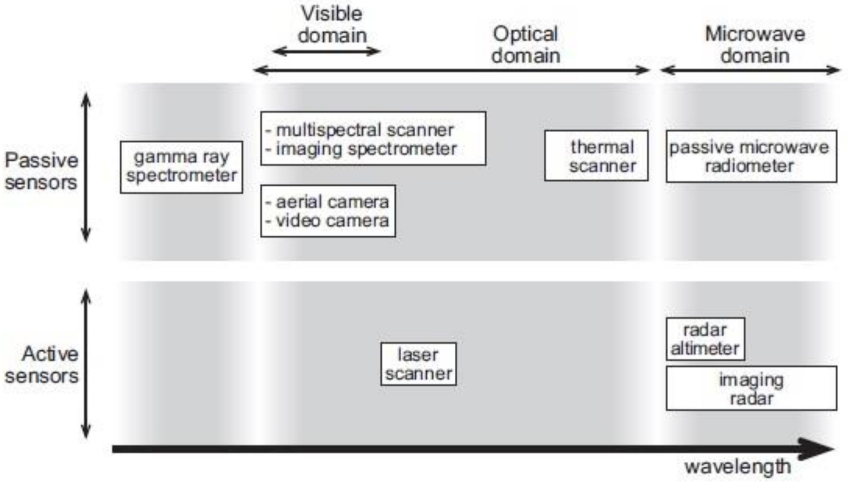


Figure (4): Overview of the sensors that are introduced in this section.

1.3.1 Passive sensors

Gamma-ray spectrometer

The gamma-ray spectrometer measures the amount of gamma rays emitted by the upper soil or rock layers due to radioactive decay. The energy measured in specific wavelength bands provides information on the abundance of (radio isotopes that relate to) specific minerals. Therefore, the main application is found in mineral exploration. Gamma rays have a very short wavelength on the order of picometres (pm)). Because of large atmospheric absorption of these waves this type of energy can only be measured up to a few hundred metres above the Earth's surface.

Aerial camera

The camera system (lens and film) is mostly found in aircraft for aerial photography.
Low orbiting satellites and NASA Space Shuttle missions also apply conventional camera techniques. The film types used in the camera enable electromagnetic energy in the range between 400 nm and 900 nm to be recorded. Aerial photographs are used in a wide range of applications. The rigid and regular geometry of aerial photographs in combination with the possibility to acquire stereo-photography has enabled the development of photogrammetric procedures' for obtaining precise 3D coordinates. Although aerial photos are used in many applications, principal applications include medium and large scale (topographic) mapping and cadastral mapping. Today, analogue photos are often scanned to be stored and processed in digital systems.

Video camera

Video cameras are sometimes used to record image data. Most video sensors are only sensitive to the visible colours, although a few are able to record the nearinfrared part of the spectrum (Figure 3.2). Until recently, only analogue video cameras were available. Today, digital video cameras are increasingly available, some of which are applied in remote sensing. Mostly, video images serve to provide low cost image data for qualitative purposes, for example, to provide additional visual information about an area captured with another sensor (e.g., laser scanner or radar).

Multispectral scanner

The multispectral scanner is an instrument that mainly measures the reflected sunlight in the optical domain. A scanner systematically 'scans' the Earth's surface thereby measuring the energy reflected from the viewed area. This is done simultaneously for several avelength bands, hence the name 'multispectral scanner'. A wavelength band is an interval of the electromagnetic spectrum for which the average reflected energy is measured. The reason for measuring a number of distinct wavelength bands is that each band is related to specific characteristics of the Earth's surface. For example, reflection characteristics of 'blue' light give information about the mineral composition; reflection characteristics of 'infrared light' tell something about the type and health of vegetation. The definition of the wavebands of a scanner, therefore, depends on the applications for which the sensor has been designed.

Imaging spectrometer

The principle of the imaging spectrometer is similar to that of the multispectral scanner, except that spectrometers measure only very narrow (5–10 nm) spectral bands. This results in an almost continuous reflectance curve per pixel rather than the values for relatively broad spectral bands. The spectral curves measured depend on the chemical composition of the material. Imaging spectrometer data, therefore, can be used to determine mineral composition of the surface or the chlorophyll content of the surface water.

Thermal scanner

Thermal scanners measure thermal data in the range of 10–14 μm. Wavelengths in this range are directly related to an objects temperature. Data on cloud, land and sea surface temperature are extremely useful for weather forecasting. For this reason, most remote sensing systems designed for meteorology include a thermal scanner. Thermal scanners can also be used to study the effects of drought ('water stress') on agricultural crops, and to monitor the temperature of cooling water discharged from thermal power plants**.**

Radiometer

EM energy with very long wavelengths (1–100 cm) is emitted from the soil and rocks on, or just below, the Earth's surface. The depth from which this energy is emitted depends on the properties, such as water content, of the specific material. Radiometers are used to detect this energy. The resulting data can be used in mineral exploration, soil mapping and soil moisture estimation

1.3.2 Active sensors

Laser scanner

Laser scanners are mounted on aircraft and use a laser beam (infrared light) to measure the distance from the aircraft to points located on the ground. This distance measurement is then combined with exact information on the aircraft's position to calculate the terrain elevation. Laser scanning is mainly used to produce detailed, high-resolution, Digital Terrain Models (DTM) for topographic mapping. Laser scanning is increasingly used for other purposes, such as the production of detailed 3D models of city buildings and for measuring tree heights in forestry.

Radar altimeter

Radar altimeters are used to measure the topographic profile parallel to the satellite orbit. They provide profiles (single lines of measurements) rather than 'image' data. Radar altimeters operate in the 1–6 cm domain and are able to determine height with a precision of 2–4 cm. Radar altimeters are useful for measuring relatively smooth surfaces such as oceans and for 'small scale' mapping of continental terrain models.

Imaging radar

Radar instruments operate in the 1–100 cm domain. As in multispectral scanning, different wavelength bands are related to particular characteristics of the Earth's surface. The radar backscatter (Figure 3.8) is influenced by the illuminating signal (microwave parameters) and the illuminated surface characteristics (orientation, roughness, di-electric constant/moisture content). Since radar is an active sensor system and the applied wavelengths are able to penetrate clouds, it has 'all-weather day-and-night' acquisition capability. The combination of two radar images of the same area can provide information about terrain heights. Combining two radar images acquired at different moments can be used to precisely assess changes in height or vertical deformations (SAR Interferometry).

1.4 Platforms

In remote sensing, the sensor is mounted on a platform. In general, remote sensing sensors are attached to moving platforms such as aircraft and satellites. Static platforms are occasionally used in an experimental context. For example, by using a multispectral sensor mounted to a pole, the changing reflectance characteristics of a specific crop during the day or season can be assessed. Airborne observations are carried out using aircraft with specific modifications to carry sensors. An aircraft that carries an aerial camera or a scanner needs a hole in the floor of the aircraft. Sometimes Ultra Light Vehicles (ULVs), balloons, Airship or kites are used for airborne remote sensing. Airborne observations are possible from 100 m up to 30–40 km height. Until recently, the navigation of an aircraft was one of the most difficult and crucial parts of airborne remote sensing. In recent years, the availability of satellite navigation technology has significantly improved the quality of flight execution. For spaceborne remote sensing, satellites are used. Satellites are launched into space with rockets. Satellites for Earth Observation are positioned in orbits between 150–36,000 km altitude. The specific orbit depends on the objectives of the mission, e.g., continuous observation of large areas or detailed observation of smaller areas.

1.4.1 Airborne remote sensing

Airborne remote sensing is carried out using different types of aircraft depending on the operational requirements and budget available. The speed of the aircraft can vary between 140–600 km/hour and is, among others, related to the mounted sensor system. Apart from the altitude, also the aircraft orientation affects the geometric characteristics of the remote sensing data acquired. The orientation of the aircraft is influenced by wind conditions and can be corrected for to some extent by the pilot. An Inertial Measurement Unit (IMU) can be installed in the aircraft to measure these rotations. Subsequently the measurements can be used to correct the sensor data for the resulting geometric distortions.

Today, most aircraft are equipped with satellite navigation technology, which yield the approximate position (RMS-error of less than 30 m). More precise positioning and navigation (up to decimetre accuracy) is possible using so-called 'differential approaches'. In this textbook we refer to satellite navigation in general, which comprises the American GPS system, the Russian Glonass system and the proposed European Galileo system.

In aerial photography the 'measurements' are 'stored' on hard-copy material: the negative film. For other sensors, e.g., a scanner, the digital data can be stored on tape or mass memory devices.

1.4.2 Spaceborne remote sensing

Spaceborne remote sensing is carried out using sensors that are mounted on satellites. The monitoring capabilities of the sensor are to a large extent determined by the parameters of the satellite's orbit. Different types of orbits are required to achieve continuous monitoring (meteorology), global mapping (land cover mapping), or selective imaging (urban areas). For remote sensing purposes, the following orbit characteristics are relevant :

• altitude, which is the distance (in km) from the satellite to the mean surface level of the Earth. Typically, remote sensing satellites orbit either at 600– 800 km (polar orbit) or at 36,000 km (geo-stationary orbit) distance from the Earth. The distance influences to a large extent which area is viewed and at which detail.

• inclination angle, which is the angle (in degrees) between the orbit and the equator. The inclination angle of the orbit determines, together with field of view of the sensor, which latitudes can be observed. If the inclination is 60^o then the satellite flies over the Earth between the latitudes 60^o South and 60^o North; it cannot observe parts of the Earth at latitudes above 60^o.

• period, which is the time (in minutes) required to complete one full orbit. A polar satellite orbits at 800 km altitude and has a period of 90 minutes. A ground speed of 28,000 km/hour is almost 8 km/s. Compare this figure with the speed of an aircraft, which is around 400 km/hour. The speed of the platform has implications for the type of images that can be acquired (time for 'exposure').

• repeat cycle, which is the time (in days) between two successive identical orbits. The revisit time, the time between two subsequent images of the same area, is determined by the repeat cycle together with the pointing capability of the sensor. Pointing capability refers to the possibility of the sensor-platform to 'look' sideways. Pushbroom scanners, such as those mounted on SPOT, IRS and IKONOS (Section 5.4), have this possibility. The following orbit types are most common for remote sensing missions:

- **Polar, or near polar, orbit.** These are orbits with inclination angle between 80 and 100 degrees and enable observation of the whole globe. The satellite is typically placed in orbit at 600–800 km altitude.
- **Sun-synchronous orbit.** An orbit chosen in such a way that the satellite always passes overhead at the same local solar time is called sun-synchronous. Most sun-synchronous orbits cross the equator at mid-morning (around 10:30 h). At that moment the Sun angle is low and the resultant shadows reveal terrain relief. Sun-synchronous orbits allow a satellite to record images at two fixed times during one 24-hour period: one during the day and one at night. Examples of near polar sun- synchronous satellites are Landsat, SPOT and IRS.
- **Geostationary orbit.** This refers to orbits in which the satellite is placed above the equator (inclination angle is 0_) at a distance of some 36,000 km. At this distance, the period of the satellite is equal to the period of the Earth. The result is that the

satellite is at a fixed position relative to the Earth. Geostationary orbits are used for meteorological and telecommunication satellites.

Today's meteorological weather satellite systems use a combination of geostationary satellites and polar orbiters. The geo-stationary satellites offer a continuous view, while the polar orbiters offer a higher resolution (Figure 4).

The data of spaceborne sensors need to be sent to the ground for further analysis and processing. Some older spaceborne systems utilized film cartridges that fell back to a designated area on Earth. In the meantime, practically all Earth Observation satellites apply satellite communication technology for downlink of the data. The acquired data are sent down to a receiving station or to another communication satellite that downlink the data to receiving antennæ on the ground. If the satellite is outside the range of a receiving station the data can be temporarily stored by a tape recorder in the satellite and transmitted later. One of the trends is that small receiving units (consisting of a small dish with a PC) are being developed for local reception of image data.

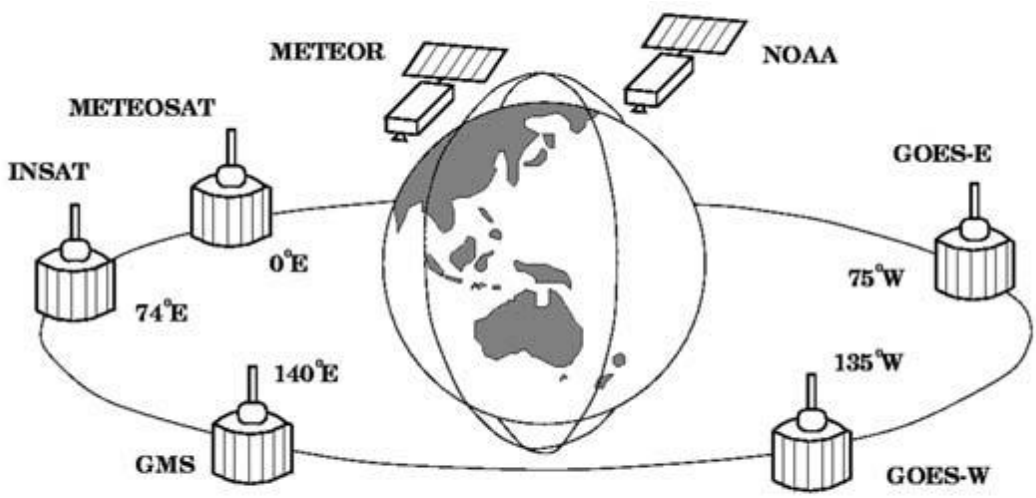
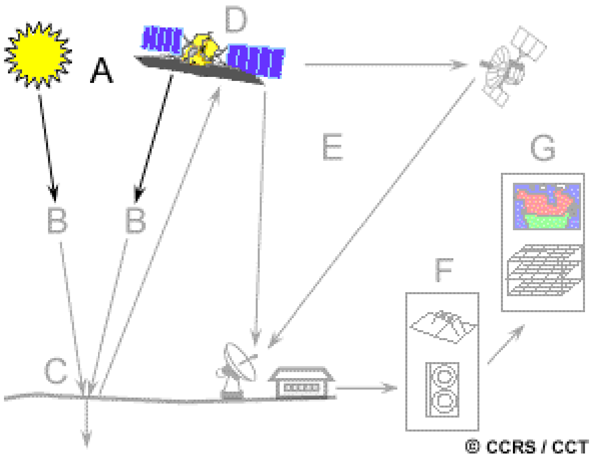


Figure (5): Meteorological observation system comprised of geo-stationary and polar satellites.

2. Electromagnetic Radiation

As was noted in the previous section, the first requirement for remote sensing is to have an **energy source to illuminate the target** (unless the sensed energy is being emitted by the target). This energy is in the form of electromagnetic radiation.



All electromagnetic radiation has fundamental properties and behaves in predictable ways according to the basics of wave theory. **Electromagnetic radiation** consists of an electrical field(E) which varies in magnitude in a direction perpendicular to the direction in which the radiation is traveling, and a magnetic field (M) oriented at right angles to the electrical field. Both these fields travel at the speed of light (c).

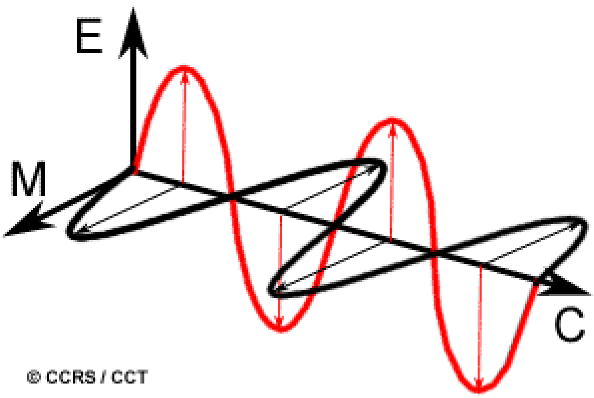


Figure (6): The Electromagnetic radiation.

Two characteristics of electromagnetic radiation are particularly important for understanding remote sensing. These are the **wavelength and frequency.**

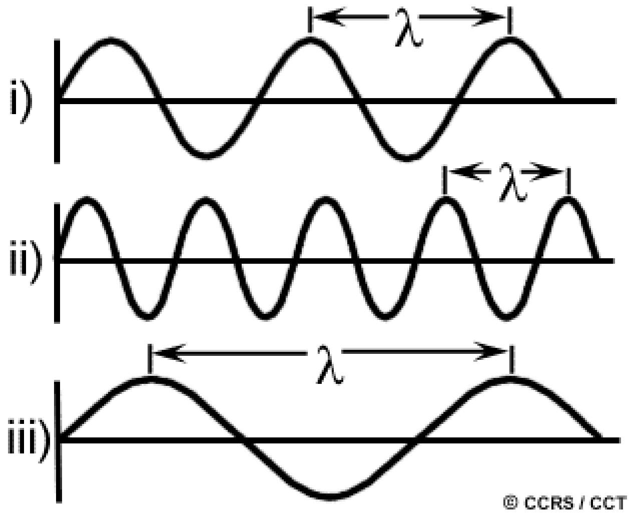


Figure (7): wavelength and frequency.

The wavelength is the length of one wave cycle, which can be measured as the distance between successive wave crests. Wavelength is usually represented by the Greek letter lambda (λ). Wavelength is measured in metres (m) or some factor of metres such as **nanometres** (nm, 10^{-9} metres), **micrometres** (μm, 10^{-6} metres) or centimetres (cm, 10^{-2} metres). Frequency refers to the number of cycles of a wave passing a fixed point per unit of time. Frequency is normally measured in **hertz** (Hz), equivalent to one cycle per second, and various multiples of hertz.

Wavelength and frequency are related by the following formula:

$$c = \lambda \nu \qquad (1)$$

Where:
λ: wavelength (m)
ν: frequency (cycles per second, Hz)
c: speed of light ($3* 10^8$ m/s)

Therefore, the two are inversely related to each other. The shorter the wavelength, the higher the frequency. The longer the wavelength, the lower the frequency. Understanding the characteristics of electromagnetic radiation in terms of their wavelength and frequency is crucial to understanding the information to be extracted from remote sensing data.

2.1 The Electromagnetic Spectrum

The **electromagnetic spectrum** ranges from the shorter wavelengths (including gamma and x-rays) to the longer wavelengths (including microwaves and broadcast radio waves). There are several regions of the electromagnetic spectrum which are useful for remote sensing.

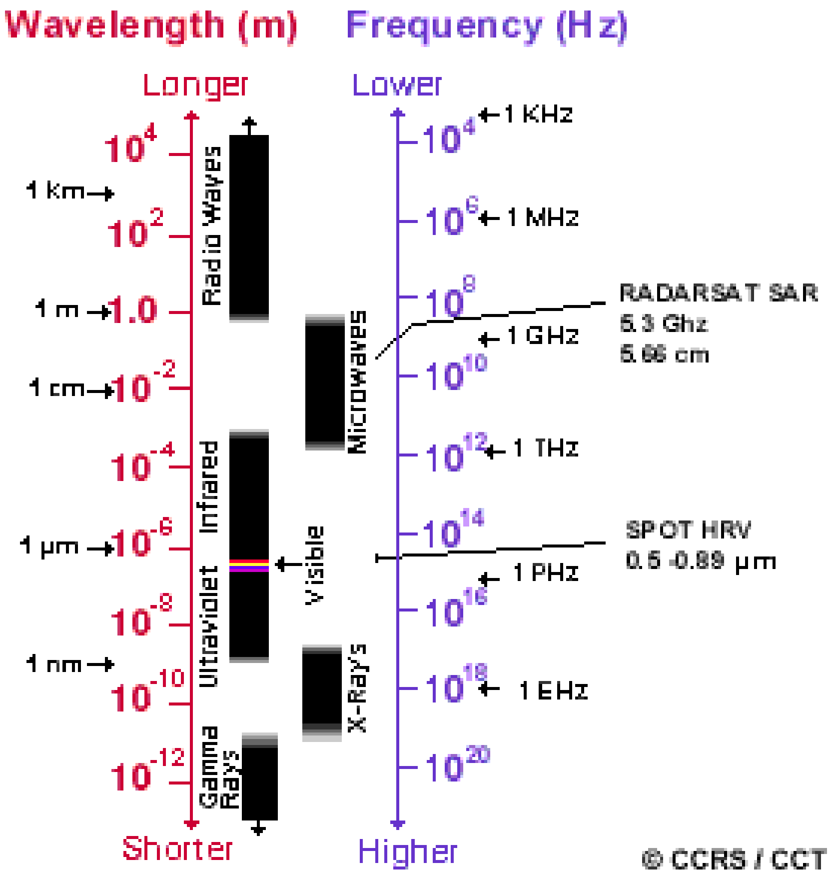
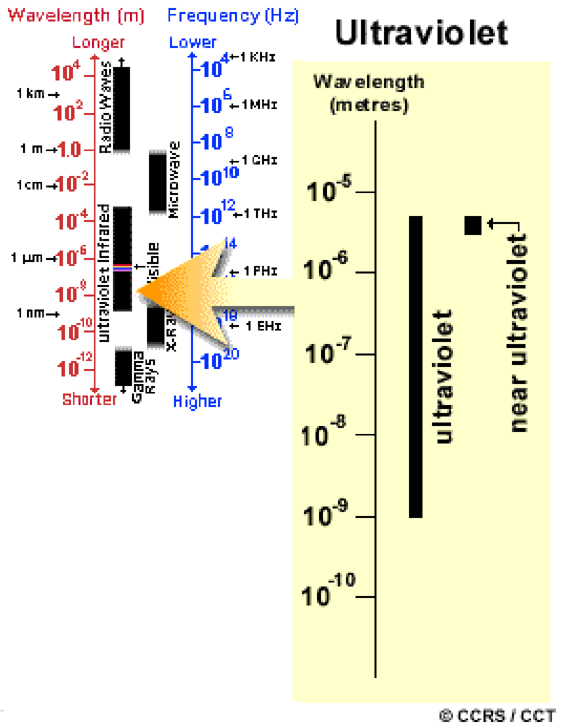
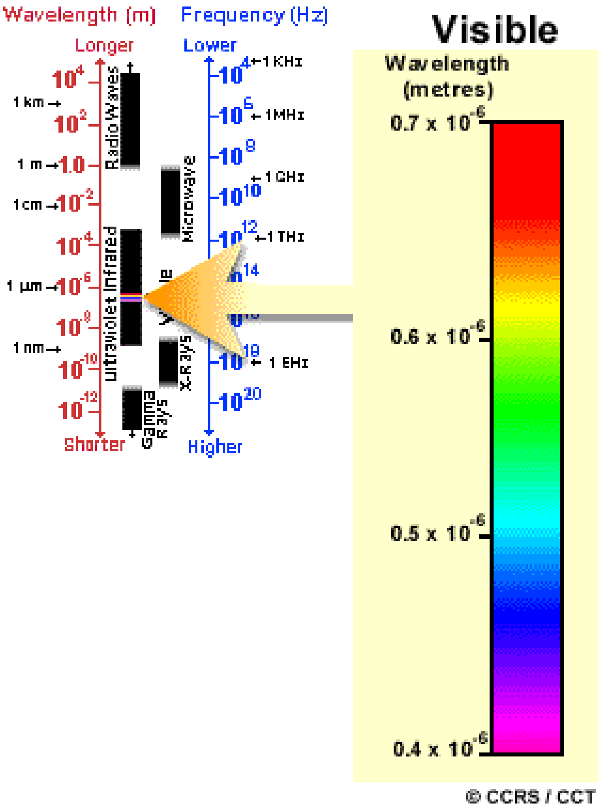


Figure (8): The Electromagnetic Spectrum.

For most purposes, the **ultraviolet or UV** portion of the spectrum has the shortest wavelengths which are practical for remote sensing. This radiation is just beyond the violet portion of the visible wavelengths, hence its name. Some Earth surface materials, primarily rocks and minerals, fluoresce or emit visible light when illuminated by UV radiation.



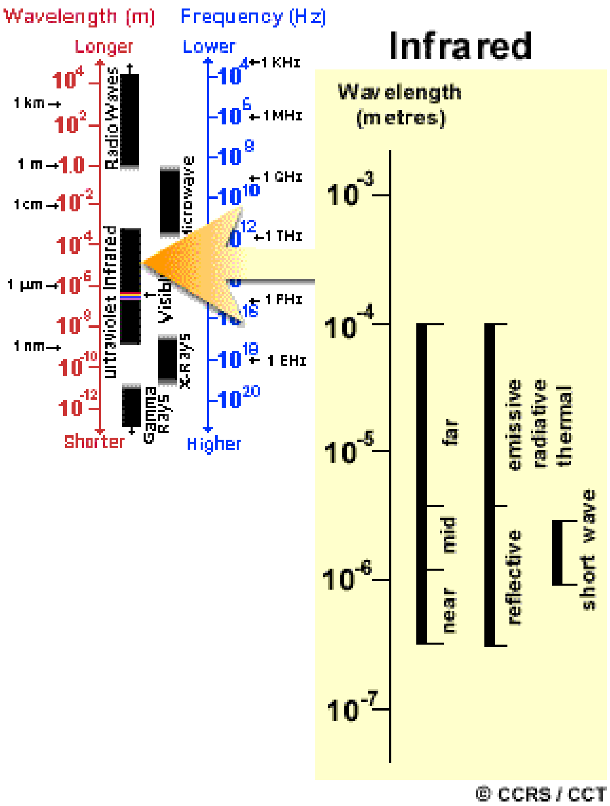
The light which our eyes - our "remote sensors" - can detect is part of the **visible spectrum**. It is important to recognize how small the visible portion is relative to the rest of the spectrum. There is a lot of radiation around us which is "invisible" to our eyes, but can be detected by other remote sensing instruments and used to our advantage. The visible wavelengths cover a range from approximately 0.4 to 0.7 µm. The longest visible wavelength is red and the shortest is violet. Common wavelengths of what we perceive as particular colours from the visible portion of the spectrum are listed below. It is important to note that this is the only portion of the spectrum we can associate with the concept of **colours**.

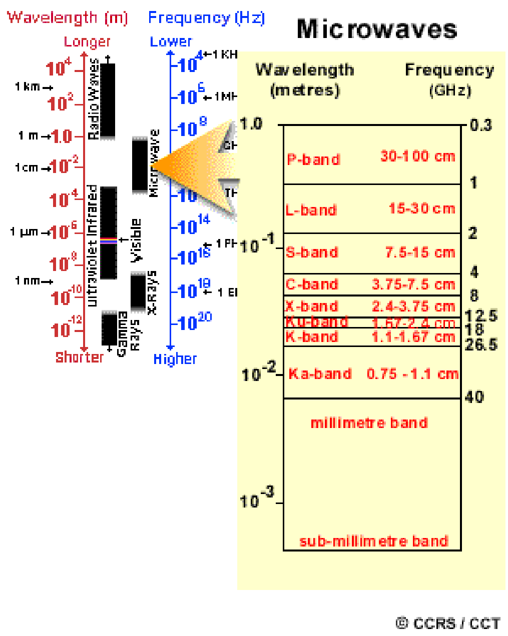


- **Violet:** 0.4 - 0.446 µm
- **Blue:** 0.446 - 0.500 µm
- **Green:** 0.500 - 0.578 µm
- **Yellow:** 0.578 - 0.592 µm
- **Orange:** 0.592 - 0.620 µm
- **Red:** 0.620 - 0.7 µm

Blue, **green**, and **red** are the **primary colours** or wavelengths of the visible spectrum. They are defined as such because no single primary colour can be created from the other two, but all other colours can be formed by combining blue, green, and red in various proportions. Although we see sunlight as a uniform or homogeneous colour, it is actually composed of various wavelengths of radiation in primarily the ultraviolet, visible and infrared portions of the spectrum. The visible portion of this radiation can be shown in its component colours when sunlight is passed through a **prism**, which bends the light in differing amounts according to wavelength.

The next portion of the spectrum of interest is the infrared (IR) region which covers the wavelength range from approximately 0.7 µm to 100 µm - more than 100 times as wide as the visible portion! The infrared region can be divided into two categories based on their radiation properties - the **reflected IR**, and the emitted or **thermal IR**. Radiation in the reflected IR region is used for remote sensing purposes in ways very similar to radiation in the visible portion. The reflected IR covers wavelengths from approximately 0.7 µm to 3.0 µm. The thermal IR region is quite different than the visible and reflected IR portions, as this energy is essentially the radiation that is emitted from the Earth's surface in the form of heat. The thermal IR covers wavelengths from approximately 3.0 µm to 100 µm. The portion of the spectrum of more recent interest to remote sensing is the **microwave region** from about 1 mm to 1 m. This covers the longest wavelengths used for remote sensing. The shorter wavelengths have properties similar to the thermal infrared region while the longer wavelengths approach the wavelengths used for radio broadcasts.



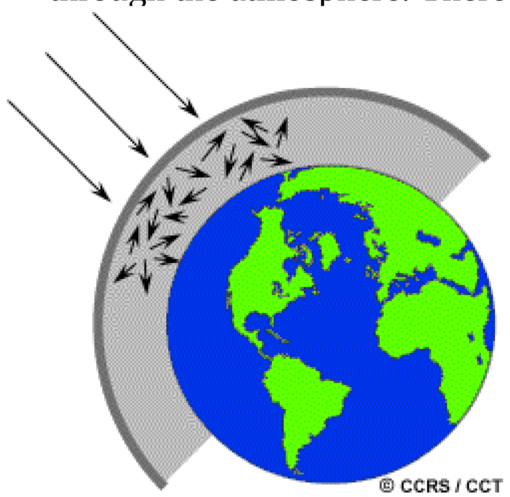


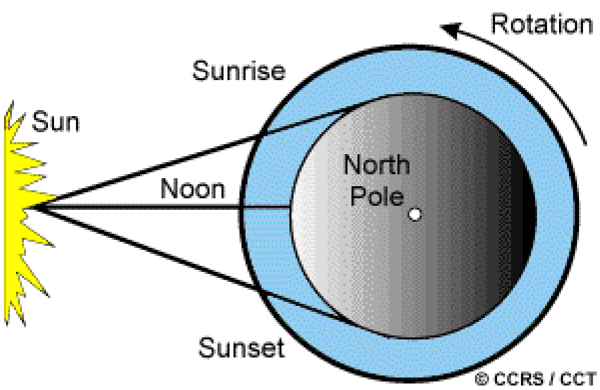
2.2 Interactions with the Atmosphere

Before radiation used for remote sensing reaches the Earth's surface it has to travel through some distance of the Earth's atmosphere. Particles and gases in the atmosphere can affect the incoming light and radiation. These effects are caused by the mechanisms of **scattering** and **absorption**.

2.2.1 Scattering.

Scattering occurs when particles or large gas molecules present in the atmosphere interact with and cause the electromagnetic radiation to be redirected from its original path. How much scattering takes place depends on several factors including the wavelength of the radiation, the abundance of particles or gases, and the distance the radiation travels through the atmosphere. There are three (3) types of scattering which take place.

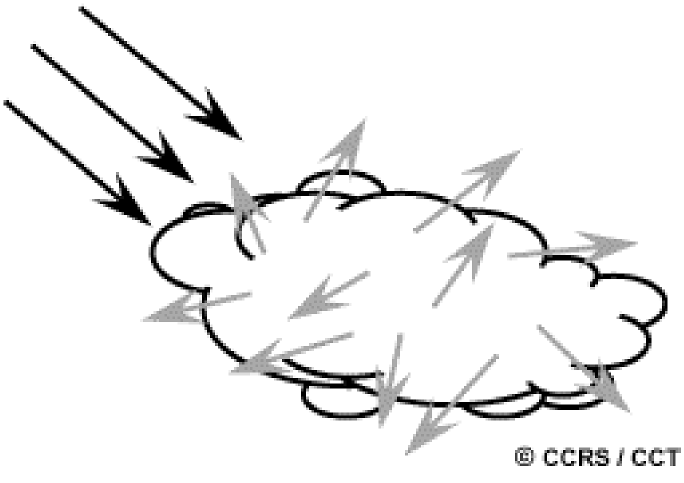




Rayleigh scattering occurs when particles are very small compared to the wavelength of the radiation. These could be particles such as small specks of dust or nitrogen and oxygen molecules. Rayleigh scattering causes shorter wavelengths of energy to be scattered much more than longer wavelengths. Rayleigh scattering is the dominant scattering mechanism in the upper atmosphere. The fact that the sky appears "blue" during the day is because of this phenomenon. As sunlight passes through the atmosphere, the shorter wavelengths (i.e. blue) of the visible spectrum are scattered more than the other (longer) visible wavelengths. At **sunrise and sunset** the light has to travel farther through the atmosphere than at midday and the scattering of the shorter wavelengths is more complete; this leaves a greater proportion of the longer wavelengths to penetrate the atmosphere.

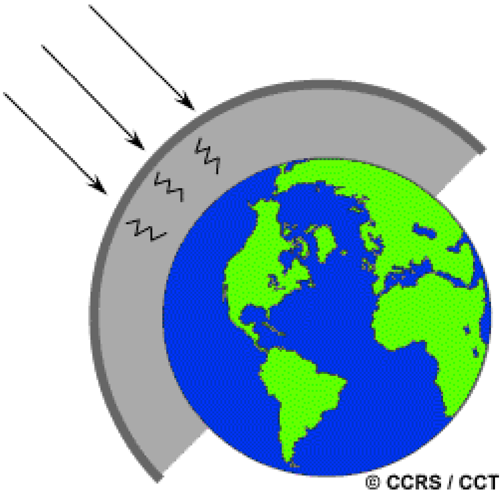
Mie scattering occurs when the particles are just about the same size as the wavelength of the radiation. Dust, pollen, smoke and water vapour are common causes of Mie scattering which tends to affect longer wavelengths than those affected by Rayleigh scattering. Mie scattering occurs mostly in the lower portions of the atmosphere where larger particles are more abundant, and dominates when cloud conditions are overcast.

The final scattering mechanism of importance is called **nonselective scattering**. This occurs when the particles are much larger than the wavelength of the radiation. Water droplets and large dust particles can cause this type of scattering. Nonselective scattering gets its name from the fact that all wavelengths are scattered about equally. This type of scattering causes fog and clouds to appear white to our eyes because blue, green, and red light are all scattered in approximately equal quantities (blue+green+red light = white light).



2.2.2 Absorption

Is the other main mechanism at work when electromagnetic radiation interacts with the atmosphere. In contrast to scattering, this phenomenon causes molecules in the atmosphere to absorb energy at various wavelengths. Ozone, carbon dioxide, and water vapour are the three main atmospheric constituents which absorb radiation. **Ozone** serves to absorb the harmful (to most living things) ultraviolet radiation from the sun. Without this protective layer in the atmosphere our skin would burn when exposed to sunlight. You may have heard **carbon dioxide** referred to as a greenhouse gas. This is because it tends to absorb radiation strongly in the far infrared portion of the spectrum - that area associated with thermal heating - which serves to trap this heat inside the atmosphere. Water vapour in the atmosphere absorbs much of the incoming longwave infrared and shortwave microwave radiation (between 22μm and 1m). The presence of water vapour in the lower atmosphere varies greatly from location to location and at different times of the year. For example, the air mass above a desert would have very little water vapour to absorb energy, while the tropics would have high concentrations of water vapour (i.e. high humidity).



Because these gases absorb electromagnetic energy in very specific regions of the spectrum, they influence where (in the spectrum) we can "look" for remote sensing urposes. Those areas of the spectrum which are not severely influenced by atmospheric absorption and thus, are useful to remote sensors, are called **atmospheric windows**. By comparing the characteristics of the two most common energy/radiation sources (the sun and the earth) with the atmospheric windows available to us, we can define those wavelengths that we can use **most effectively** for remote sensing. The visible portion of the spectrum, to which our eyes are most sensitive, corresponds to both an atmospheric window and the peak energy level of the sun. Note also that heat energy emitted by the Earth corresponds to a window around 10 μm in the thermal IR portion of the spectrum, while the large window at wavelengths beyond 1 mm is associated with the microwave region.

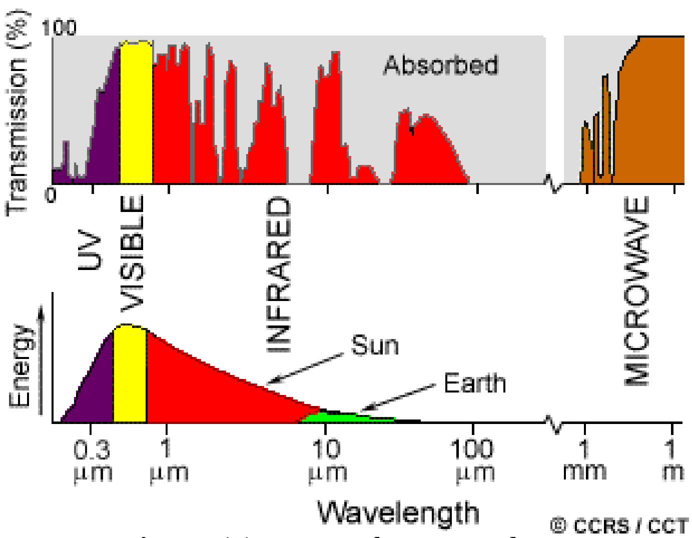
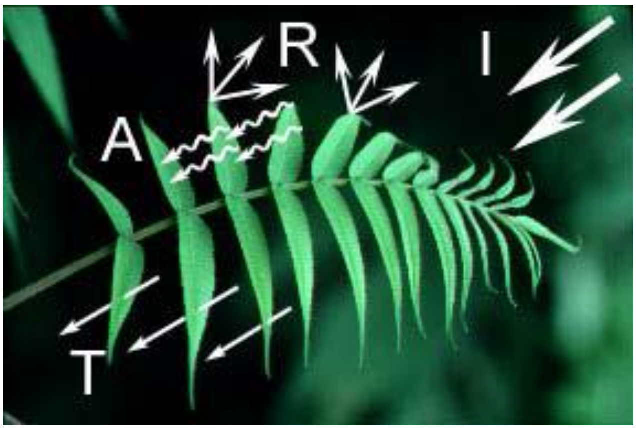


Figure (9): Atmospheric Windows

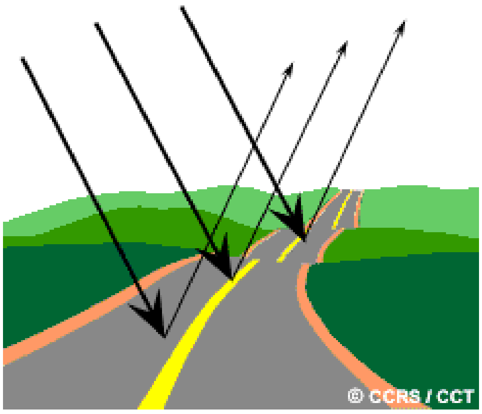
Now that we understand how electromagnetic energy makes its journey from its source to the surface (and it is a difficult journey, as you can see) we will next examine what happens to that radiation when it does arrive at the Earth's surface.

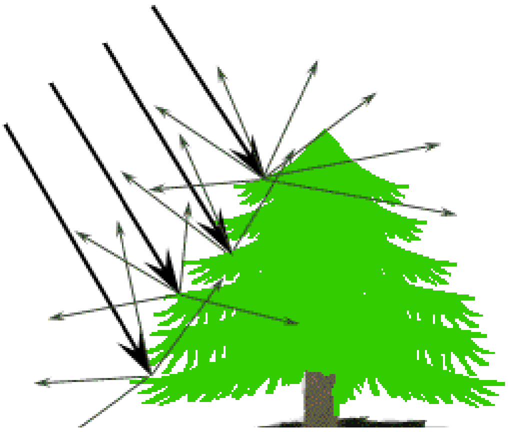
2.3 Radiation - Target Interactions

Radiation that is not absorbed or scattered in the atmosphere can reach and interact with the Earth's surface. There are three (3) forms of interaction that can take place when energy strikes, or is **incident (I)** upon the surface. These are: **absorption (A)**; **transmission (T)**; and **reflection (R)**. The total incident energy will interact with the surface in one or more of these three ways. The proportions of each will depend on the wavelength of the energy and the material and condition of the feature.



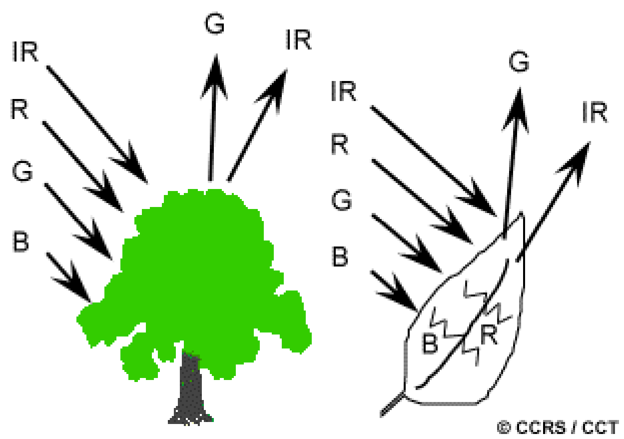
Absorption (A) occurs when radiation (energy) is absorbed into the target while transmission (T) occurs when radiation passes through a target. Reflection (R) occurs when radiation "bounces" off the target and is redirected. In remote sensing, we are most interested in measuring the radiation reflected from targets. We refer to two types of reflection, which represent the two extreme ends of the way in which energy is reflected from a target: **specular reflection** and **diffuse reflection**.





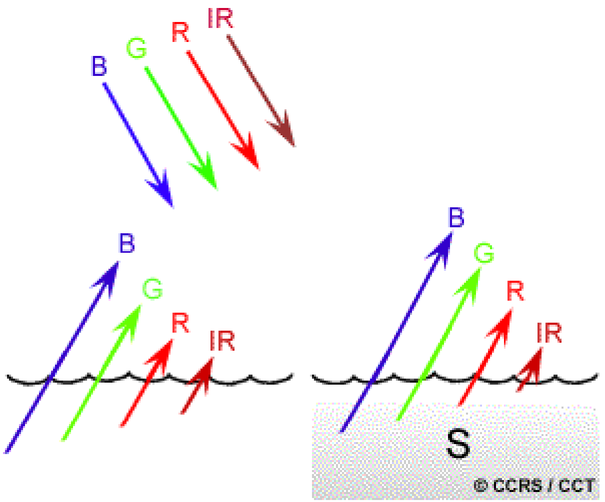
When a surface is smooth we get **specular** or mirror-like reflection where all (or almost all) of the energy is directed away from the surface in a single direction. **Diffuse** reflection occurs when the surface is rough and the energy is reflected almost uniformly in all directions. Most earth surface features lie somewhere between perfectly specular or perfectly diffuse reflectors. Whether a particular target reflects specularly or diffusely, or somewhere in between, depends on the surface roughness of the feature in comparison to the wavelength of the incoming radiation. If the wavelengths are much smaller than the surface variations or the particle sizes that make up the surface, diffuse reflection will dominate. For example, finegrained sand would appear fairly smooth to long wavelength microwaves but would appear quite rough to the visible wavelengths. Let's take a look at a couple of examples of targets at the Earth's surface and how energy at the visible and infrared wavelengths interacts with them.

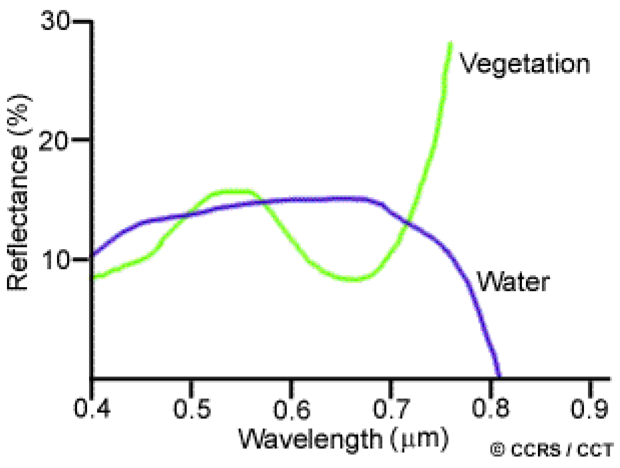
Leaves: A chemical compound in leaves called chlorophyll strongly absorbs radiation in the red and blue wavelengths but reflects green wavelengths. Leaves appear "greenest" to us in the summer, when chlorophyll content is at its maximum. In autumn, there is less chlorophyll in the leaves, so there is less absorption and proportionately more reflection of the red wavelengths, making the leaves appear red or yellow (yellow is a combination of red and green wavelengths). The internal structure of healthy leaves act as excellent diffuse reflectors of near-infrared wavelengths. If our eyes were sensitive to near-infrared, trees would appear extremely bright to us at these wavelengths. In fact, measuring and monitoring the near-IR reflectance is one way that scientists can determine how healthy (or unhealthy) vegetation may be.



Water: Longer wavelength visible and near infrared radiation is absorbed more by water than shorter visible wavelengths. Thus water typically looks blue or blue-green due to stronger reflectance at these shorter wavelengths, and darker if viewed at red or near infrared wavelengths. If there is suspended sediment present in the upper layers of the water body, then this will allow better reflectivity and a brighter appearance of the water. The apparent colour of the water will show a slight shift to longer
wavelengths. Suspended sediment (S) can be easily confused with shallow (but clear) water, since these two phenomena appear very similar. Chlorophyll in algae absorbs more of the blue wavelengths and reflects the green, making the water appear more green in colour when algae is present. The topography of the water surface (rough, smooth, floating

materials, etc.) can also lead to complications for water-related interpretation due to potential problems of specular reflection and other influences on colour and brightness.
We





We can see from these examples that, depending on the complex make-up of the target that is being looked at, and the wavelengths of radiation involved, we can observe very different responses to the mechanisms of absorption, transmission, and reflection. By measuring the energy that is reflected (or emitted) by targets on the Earth's surface over a variety of different wavelengths, we can build up a **spectral response** for that object. By comparing the response patterns of different features we may be able to distinguish between them, where we might not be able to, if we only compared them at one wavelength. For example, water and vegetation may reflect somewhat similarly in the visible wavelengths but
are almost always separable in the infrared. Spectral response can be quite variable, even for the same target type, and can also vary with time (e.g. "green-ness" of leaves) and location. Knowing where to "look" spectrally and understanding the factors which influence the spectral response of the features of interest are critical to correctly interpreting the interaction of electromagnetic radiation with the surface.

3. Electromagnetic Wave Propagation

The thermal and compositional state of the atmosphere affects both the generation and propagation of electromagnetic (EM) waves. For now, we ignore the source of the EM waves and focus instead on their propagation through a homogeneous, lossless medium.

2.1 Maxwell's Equations and the Wave Equation

The behavior of electromagnetic waves in free space is governed by Maxwell's equations:

$$\nabla \times E = -\frac{\partial B}{\partial t} \quad (2)$$

$$\nabla \times H = \frac{\partial D}{\partial t} + J \quad (3)$$

$$B = \mu_0 \mu_r H \quad (4)$$

$$D = \varepsilon_0 \varepsilon_0 E \quad (5)$$

$$\nabla . \vec{E} = 0 \quad (6)$$

$$\nabla . \vec{H} = 0 \quad (7)$$

where
E = electric vector
D = displacement vector
H = magnetic vector
B = induction vector
$\mu_0 \varepsilon_0$: permeability and permittivity of vacuum
$\mu_r \varepsilon_r$: relative permeability and permittivity

Maxwell's concept of electromagnetic waves is that a smooth wave motion exists in the magnetic and electric force fields. In any region in which there is a temporal change in the electric field, a magnetic field appears automatically in that same region as a conjugal partner and vice versa. This is expressed by the above coupled equations.

3.2 Wave Equation and Solution

In homogeneous, isotropic, and nonmagnetic media, Maxwell's equations can be combined to derive the wave equation:

$$\nabla^2 E = \mu_0 \varepsilon_0 \mu_r \varepsilon_r \frac{\partial^2 E}{\partial t^2} \quad (8)$$

or, in the case of a sinusoidal field,

$$\nabla^2 E + \frac{\omega^2}{{c_r}^2} E = 0 \qquad (9)$$

Where

$$c_r = \frac{1}{\sqrt{\mu_0 \varepsilon_0 \mu_r \varepsilon_r}} = \frac{c}{\sqrt{\mu_r \varepsilon_r}} \qquad (10)$$

Usually, μ_r = 1 and ε_r varies from 1 to 80 and is a function of the frequency. The solution for the above differential equation is given by

$$E = Ae^{i(kr - \omega t + \emptyset)} \qquad (11)$$

where A is the wave amplitude, ω is the angular frequency, $\emptyset$ is the phase, and k is the wave vector in the propagation medium ($k = 2\pi\sqrt{\varepsilon_r}/\lambda$, λ = wavelength $= \frac{2\pi c}{\omega}$, c = speed of lightin vacume) The wave frequency v is defined as $v = \omega/2\pi$.
Remote sensing instruments exploit different aspects of the solution to the wave equation in order to learn more about the properties of the medium from which the radiation is being sensed. For example, the interaction of electromagnetic waves with natural surfaces and atmospheres is strongly dependent on the frequency of the waves. This will manifest itself in changes in the amplitude [the magnitude of **A** in Equation (11)] of the received wave as the frequency of the observation is changed. This type of information is recorded by multispectral instruments such as the Landsat Thematic Mapper and the Advanced Advanced Spaceborne Thermal Emission and Reflection Radiometer.
In other cases, one can infer information about the electrical properties and geometry of the surface by observing the polarization [the vector components of A in Equation (11) of the received waves. This type of information is recorded by polarimeters and polarimetric radars. Doppler lidars and radars, on the other hand, measure the change in frequency between the transmitted and received waves in order to infer the velocity with which an object or medium is moving. This information is contained in the angular frequency ω of the wave shown in Equation (11). The quantity $kr - \omega t + \emptyset$ in Equation (11) is known as the phase of the wave. This phase changes by 2 π every time the wave moves through a distance equal to the wavelength λ. Measuring the phase of a wave therefore provides an extremely accurate way to measure the distance that the wave actually travelled. Interferometers exploit this property of the wave to accurately measure differences in the path length between a source and two collectors, allowing one to significantly increase the resolution with which the position of the source can be established.

3.3 Quantum Properties of Electromagnetic Radiation

Maxwell's formulation of electromagnetic radiation leads to a mathematically smooth wave motion of fields. However, at very short wavelengths, it fails to account for certain significant phenomena that occur when the wave interacts with matter. In this case, a quantum description is more appropriate. The electromagnetic energy can be presented in a quantized form as bursts of radiation with a quantized radiant energy Q, which is proportional to the frequency *v*:

$$Q = h\nu \tag{12}$$

where h = Planck's constant = 6.626 × 10–34 joule second. The radiant energy carried by the wave is not delivered to a receiver as if it is spread evenly over the wave, as Maxwell had visualized, but is delivered on a probabilistic basis. The probability that a wave train will make full delivery of its radiant energy at some place along the wave is proportional to the flux density of the wave at that place. If a very large number of wave trains are coexistent, then the overall average effect follows Maxwell's equations.

3.4 Polarization

As mentioned previously, electromagnetic radiation consists of electric and magnetic fields which oscillate with the frequency of radiation. These fields are always perpendicular to each other. So, it is possible to specify the orientation of the electromagnetic radiation by specifying the orientation of one of those fields. The orientation of the electric field is defined as the orientation of electromagnetic field and this is called as **.polarization..** In other words, polarization refers to the orientation of the electrical field component of an electromagnetic wave.

The plane of polarization contains both the electric vector and the direction of propagation. Simply because the plane is two-dimensional, the electric vector in the plane at a point in space can be decomposed into two orthogonal components. Call these the **x** and **y** components (following the conventions of analytic geometry). For a simple harmonic wave, where the amplitude of the electric vector varies in a sinusoidal manner, the two components have exactly the same frequency. However, these components have two other defining characteristics that can differ. First, the two components may not have the same amplitude. Second, the two components may not have the same phase, so they may not reach their maxima and minima at the same time in the fixed plane we are talking about. By considering the shape traced out in a fixed plane by the electric vector as such a plane wave passes over it, we obtain a description of the **polarization state**. By considering that issue, three types of polarization of electromagnetic waves can be defined. These are Linear, Circular and Elliptical Polarizations.

a) Linear polarization

If the electrical vector remains in one plane, then the wave is linearly polarised. By convention, if the electric vector (or field) is parallel to the earth's surface, the wave is said to be horizontally polarized, if the electric vector (or field) is perpendicular to the earth's surface, the wave is said to be vertically polarized.

Linear polarization is shown in Figure 10. Here, two orthogonal components **(**Ex-red and

Ey-green**)** of the electric field vector **(**E-blue**)** are in phase and they form a path **(**purple**)** in the plane while propagating. In linear polarization case, the strength of the two components are always equal or related by a constant ratio, so the direction of the electric vector (the vector sum of these two components) will always fall on a single line in the plane. We call this special case linear polarization. The direction of this line will depend on the relative amplitude of the two components. This direction can be in any angle in the plane but the direction never varies.

Linear polarization is most often used in conventional radar antennas since it is the easiest to achieve. The choice between horizontal and vertical polarisation is often left to the discretion of the antenna designer, although the radar system engineer might sometimes want to specify one or the other, depending upon the importance of ground reflections.

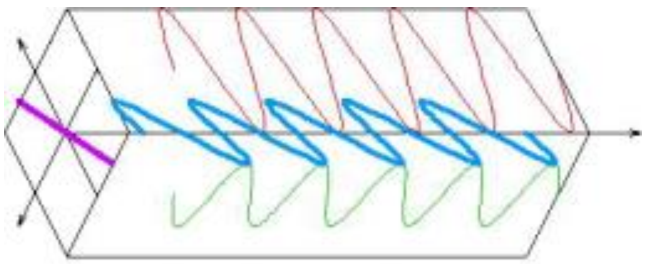


Figure 10: Linear Polarization.

b) Circular polarization

In case of circular polarization as shown in Figure 11, the two orthogonal components (Exred and Ey-green) of the electric field vector (E-blue) have exactly the same amplitude and are exactly ninety degrees out of phase. They form a path (purple) in the plane while propagating. In this case, one component is zero when the other component is at maximum or minimum amplitude. Notice that there are two possible phase relationships that satisfy this requirement. The **x** component can be ninety degrees ahead of the **y** component or it can be ninety degrees behind the **y** component. In this special case the electrical vector will be rotating in a circle while the wave propagates. The direction of rotation will depend on which of the two phase relationships exists. One rotation is finished after one wavelength. The rotation may be left or right handed. In another word, image of the electric field vector (E) will be circular and electromagnetic wave will be circularly polarized. Circular polarization is often desirable to attenuate the reflections of rain with respect to aircraft.

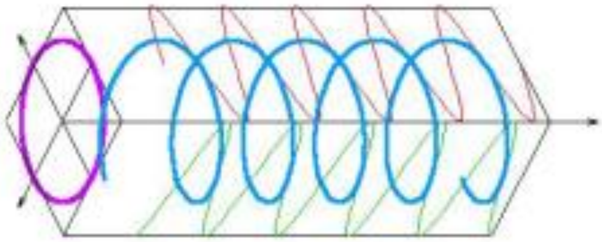


Figure 11: Circular Polarization.

c) Elliptical polarization

As shown in Figure 12, two components (Ex-red and Ey-green) of the electric field vector (E-blue) are not in phase and do not have the same amplitude and/or are not ninety degrees out of phase either. So the path (purple) formed in the plane while propagating will trace out an ellipse and this is called as elliptical polarization. In other words, image of electric field vector E will be elliptical and electromagnetic wave will be elliptically polarized. In fact linear and circular polarizations are special cases of the elliptical polarization.

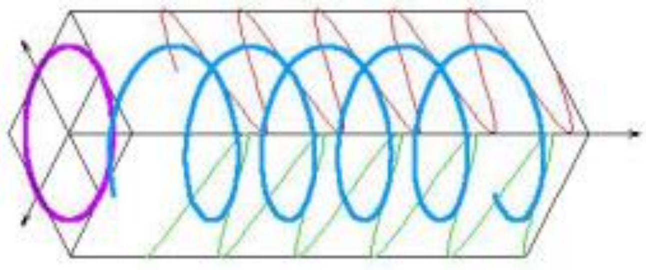


Figure (12): Elliptical polarization

3.5 Coherency

In the case of a monochromatic wave of certain frequency v_0, the instantaneous field at any point P is well defined. If the wave consists of a large number of monochromatic waves with frequencies over a bandwidth ranging from v_0 to $v_0 + \Delta v$, then the random addition of all the component waves will lead to irregular fluctuations of the resultant field. The coherency time Δt is defined as the period over which there is strong correlation of the field amplitude. More specifically, it is the time after which two waves at v and $v + \Delta v$ are out of phase by one cycle; that is, it is given by

$$v\Delta t + 1 = (v + \Delta v)\Delta t \rightarrow \Delta v\Delta t = 1 \qquad (12)$$

$$\rightarrow \Delta t = \frac{1}{\Delta v}$$

The coherence length is defined as

$$\Delta l = c\Delta t = \frac{c}{\Delta v} \qquad (13)$$

Two waves or two sources are said to be coherent with each other if there is a systematic relationship between their instantaneous amplitudes. The amplitude of the resultant field varies between the sum and the difference of the two amplitudes. If the two waves are incoherent, then the power of the resultant wave is equal to the sum of the power of the two constituent waves. Mathematically, let $E_1(t)$ and $E_2(t)$ be the two component fields at a certain location. Then the total field is

$$E(t) = E_1(t) + E_2(t) \tag{14}$$

The average power is

$$P \sim \overline{[E(t)]^2} = \overline{[E_1(t) + E_2(t)]^2} \tag{15}$$

$$= \overline{E_1(t)^2} + \overline{E_1(t)^2} + \overline{2E_1(t)E_2(t)}$$

If the two waves are incoherent relative to each other, then $\overline{E_1(t)E_2(t)} = 0$ and $P = P_1 + P_2$. If the waves are coherent, then $\overline{E_1(t)E_2(t)} \neq 0$. In the latter case, we have

$$\begin{array}{ll} P > P_1 + P_2 & \text{in some locations} \\ P < P_1 + P_2 & \text{in other locations} \end{array}$$

This is the case of optical interference fringes generated by two overlapping coherent optical beams. The bright bands correspond to where the energy is above the mean and the dark bands correspond to where the energy is below the mean.

3.6 Group and Phase Velocity

The phase velocity is the velocity at which a constant phase front progresses (see Fig13)

$$v_p = \frac{\omega}{k} \tag{16}$$

If we have two waves characterized by $(\omega - \Delta\omega, k - \Delta k)$and $(\omega + \Delta\omega, k + \Delta k)$, then the total wave is given by

$$E(z,t) = Ae^{i[(k-\Delta k)z-(\omega-\Delta\omega)t]} + Ae^{i[(k+\Delta k)z-(\omega+\Delta\omega)t]} \tag{17}$$

$$= 2Ae^{1(kz-\omega t)} cos(\Delta kz - \Delta\omega t)$$

In this case, the plane of constant amplitude moves at a velocity v_g, called the group velocity:

$$v_g = \frac{\Delta\omega}{\Delta k} \tag{18}$$

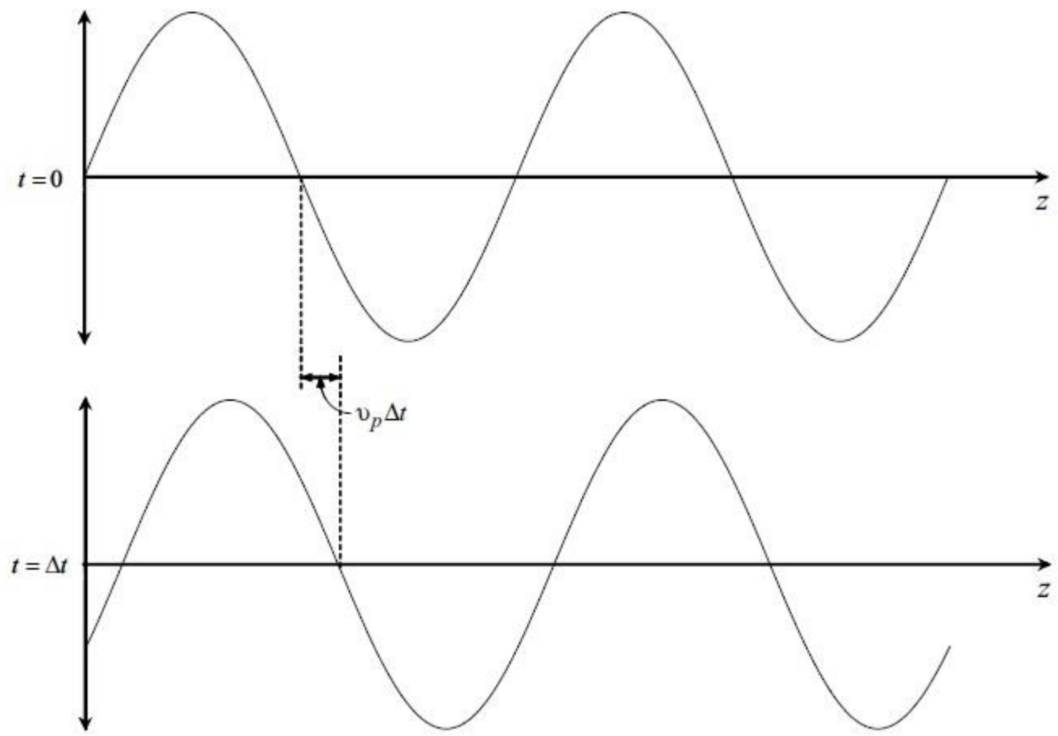


Figure (13): Phase velocity.

This is illustrated in Figure 14. It is important to note that v_g represents the velocity of propagation of the wave energy. Thus, the group velocity v_g must be equal to or smaller than the speed of light *c*. However, the phase velocity v_p can be larger than *c*. If the edium is nondispersive, then

$$\omega = ck \tag{19}$$

This implies that

$$v_p = \frac{\omega}{k} = c \tag{20}$$

And

$$v_g = \frac{\Delta\omega}{\Delta k} = c \tag{21}$$

if the medium is dispersive (i.e., ω is a nonlinear function of k), such as in the case of ionospheres, then the two velocities are different.

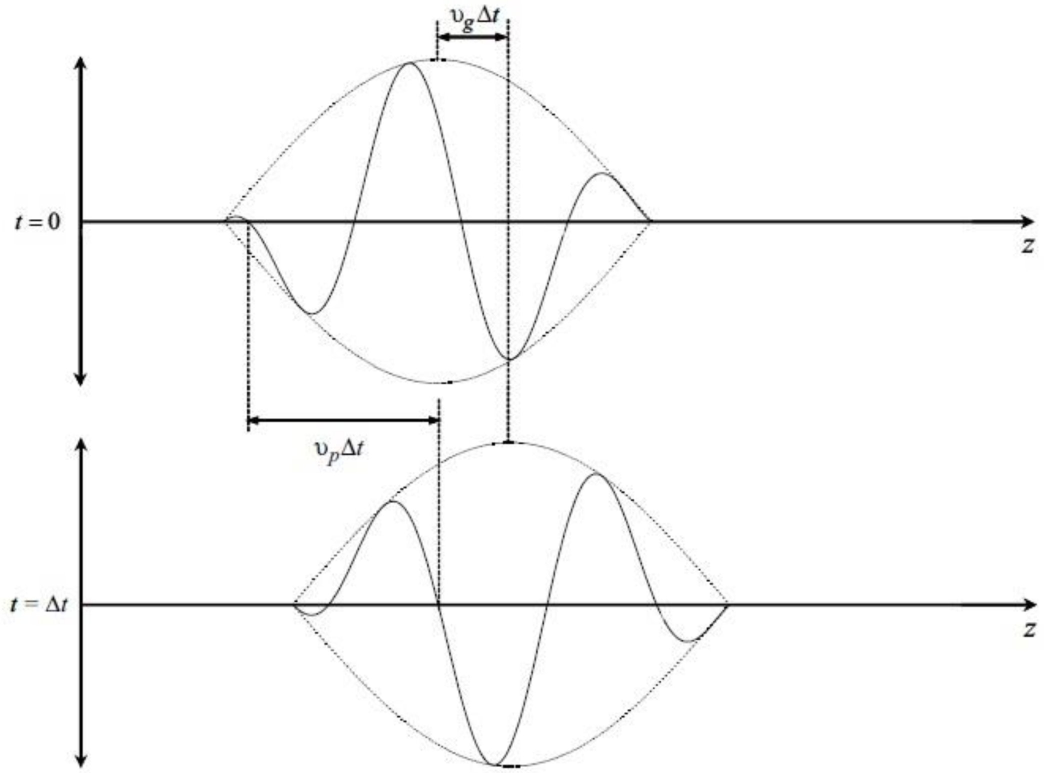


Figure 14: Group velocity.

3.7 Doppler Effect

If the relative distance between a source radiating at a fixed frequency ν and an observer varies, the signal received by the observer will have a frequency $\acute{\nu}$, which is different than ν. The difference, $\nu_d = \acute{\nu} - \nu$, is called the Doppler shift. If the source–observer distance is decreasing, the frequency received is higher than the frequency transmitted, leading to a positive Doppler shift ($\nu_d > 0$). If the source–observer distance is increasing, the reverse effect occurs (i.e., $\nu_d < 0$) and the Doppler shift is negative. The relationship between ν_d and ν is

$$\nu_d = \nu \frac{v}{c} cos\theta \qquad (22)$$

where v is the relative speed between the source and the observer, c is the velocity of light, and $\theta$is the angle between the direction of motion and the line connecting the source and the observer (see Fig. 15). The above expression assumes no relativistic effects ($v \ll c$), and it can be derived in the following simple way.

Referring to Figure16, assume an observer is moving at a velocity v with an angle θrelative to the line of propagation of the wave. The lines of constant wave amplitude are separated by the distance λ (i.e., wavelength) and are moving at velocity c. For the observer, the apparent frequency $\acute{v}$ is equal to the inverse of the time period T' that it takes the observer to cross two successive equiamplitude lines. This is given by the expression

$$cT' + vT'cos\theta = \lambda \qquad (23)$$

which can be written as

$$\frac{c}{\acute{v}} + \frac{\mathrm{v}cos\theta}{\acute{v}} = \frac{c}{v} \Rightarrow \acute{v} = v + v\frac{\mathrm{v}}{c}cos\theta = v + v_d \qquad (24)$$

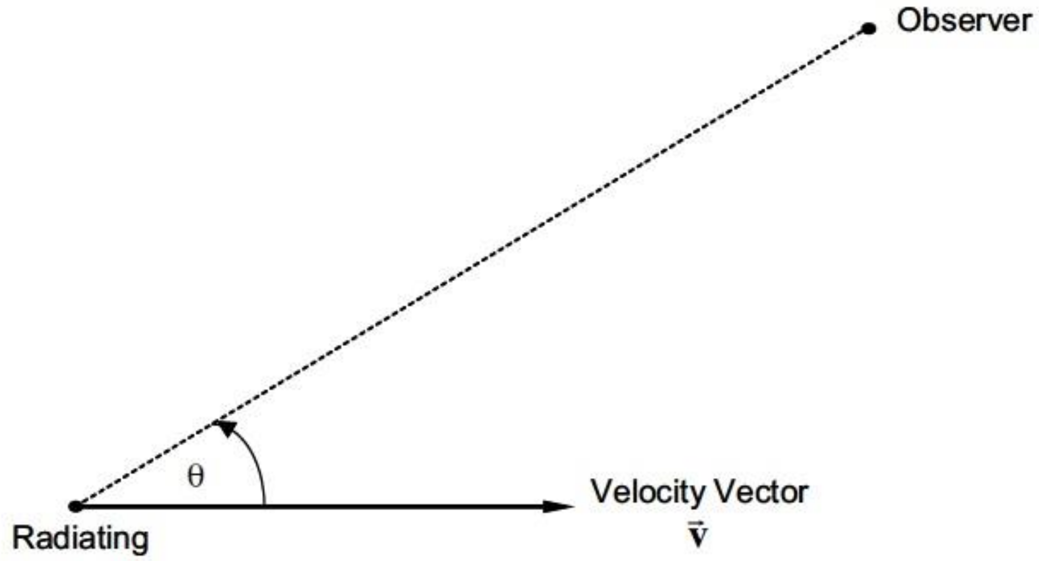


Figure (16): Doppler geometry for a moving source, fixed observer.

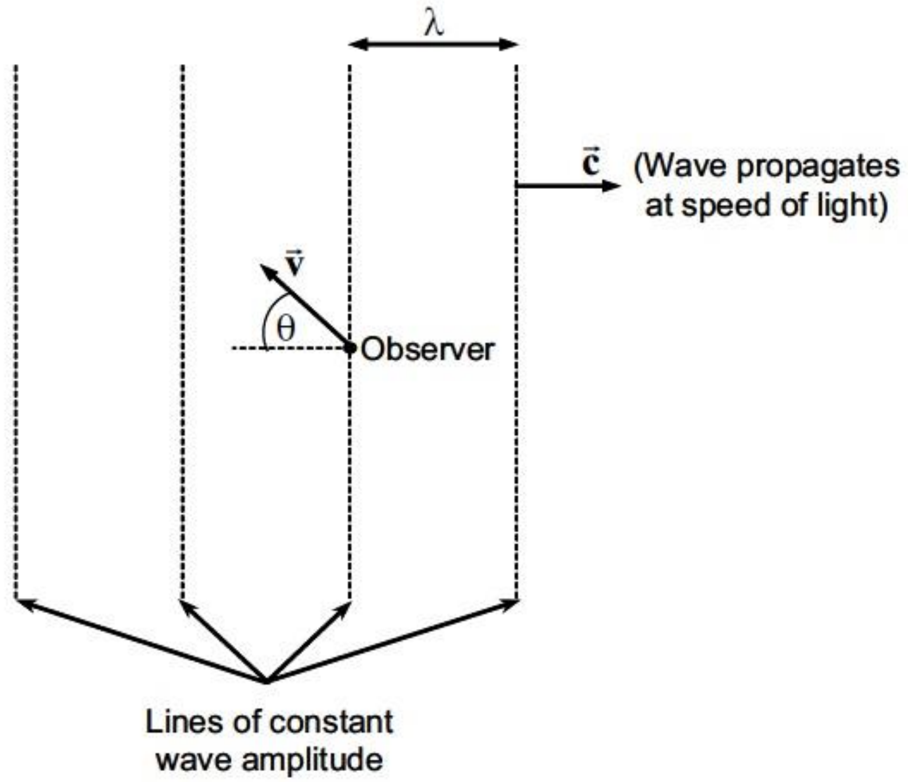


Figure (17): Geometry illustrating wave fronts passing by a moving observer.

The Doppler effect also occurs when the source and observer are fixed relative to each other but the scattering or reflecting object is moving (see Fig. 18). In this case, the Doppler shift is given by

$$v_d = v\frac{\nu}{c}(cos\theta_1 + cos\theta_2) \qquad (25)$$

and if the source and observer are collocated (i.e., θ_1 = θ_2 = θ), then

$$v_d = 2v\frac{\nu}{c}cos\theta \qquad (26)$$

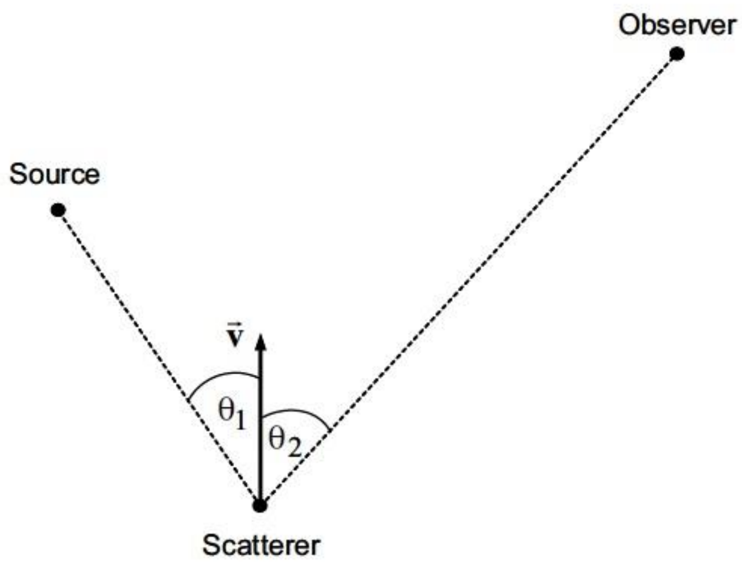


Figure (18): Doppler geometry for a moving scatterer with fixed source and observer.

The Doppler effect is used in remote sensing to measure target motion. It is also the basic physical effect used in synthetic-aperture imaging radars to achieve very high resolution imaging.

4. Radiation Quantities

A number of quantities are commonly used to characterize the electromagnetic radiation and its interaction with matter. These are briefly described below and summarized in Table 1.

Radiant energy. The energy carried by an electromagnetic wave. It is a measure of the capacity of the wave to do work by moving an object by force, heating it, or changing its state. The amount of energy per unit volume is called radiant energy density.

Radiant flux. The time rate at which radiant energy passes a certain location. It is closely related to the wave *power,* which refers to the time rate of doing work. The term *flux* is also used to describe the time rate of flow of quantized energy elements such as photons. Then the term *photon flux* is used.

Radiant flux density. Corresponds to the radiant flux intercepted by a unit area of a plane surface. The density for flux incident upon a surface is called irradiance. The density for flux leaving a surface is called exitance or emittance.

Solid angle. The solid angle Ω subtended by area A on a spherical surface is equal to the area A divided by the square of the radius of the sphere.

Radiant intensity. The radiant intensity of a point source in a given direction is the radiant flux per unit solid angle leaving the source in that direction.

Radiance. The radiant flux per unit solid angle leaving an extended source in a given direction per unit projected area in that direction (see Fig. 19). If the radiance does not change as a function of the direction of emission, the source is called Lambertian. A piece of white matte paper, illuminated by diffuse skylight, is a good example of a Lambertian source.

Hemispherical reflectance. The ratio of the reflected exitance (or emittance) from a plane of material to the irradiance on that plane.

Hemispherical transmittance. The ratio of the transmitted exitance, leaving the opposite side of the plane, to the irradiance.

Table (1): Radiation Quantities

Quantity	Usual symbol	Defining equation	Units
Radiant energy	Q		joule
Radiant energy density	W	$W=\frac{dQ}{dV}$	joule/m^3
Radiant flux	Φ	$\Phi=\frac{dQ}{dt}$	watt
Radiant flux density	E (irradiance) M (emittance)	$E, M=\frac{d\Phi}{dA}$	watt/m^2
Radiant intensity	I	$I=\frac{d\Phi}{d\Omega}$	watt/steradian
Radiance	L	$L=\frac{dI}{dA\cos\theta}$	watt/steradian m^2
Hemispherical reflectance	ρ	$\rho=\frac{M_r}{E}$	
Hemispherical absorptance	α	$\alpha=\frac{M_a}{E}$	
Hemispherical transmittance	τ	$\tau=\frac{M_t}{E}$	

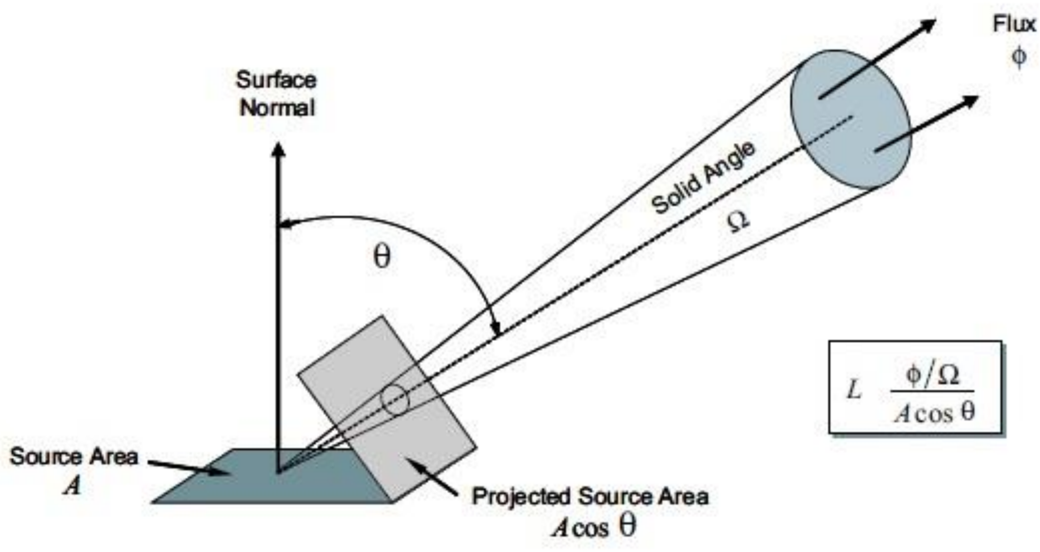


Figure (19): Concept of radiance.

4- Detection of Electromagnetic Radiation

The radiation emitted, reflected, or scattered from a body generates a radiant flux density in the surrounding space that contains information about the body's properties. To measure the properties of this radiation, a collector is used, followed by a detector. The collector is a collecting aperture that intercepts part of the radiated field. In the microwave region, an antenna is used to intercept some of the electromagnetic energy. Examples of antennas include dipoles, an array of dipoles, or dishes. In the case of dipoles, the surrounding field generates a current in the dipole with an intensity proportional to the field intensity and a frequency equal to the field frequency. In the case of a dish, the energy collected is usually focused onto a limited area where the detector (or waveguide connected to the detector) is located.

In the IR, visible, and UV regions, the collector is usually a lens or a reflecting surface that focuses the intercepted energy onto the detector. Detection then occurs by transforming the electromagnetic energy into another form of energy such as heat, electric current, or state change. Depending on the type of the sensor, different properties of the field are measured. In the case of synthetic-aperture imaging radars, the amplitude, polarization, frequency, and phase of the fields are measured at successive locations along the flight line. In the case of optical spectrometers, the energy of the field at a specific location is measured as a function of wavelength.

In the case of radiometers, the main parameter of interest is the total radiant energy flux. In the case of polarimeters, the energy flux at different polarizations of the wave vector is measured. In the case of x-ray and gamma-ray detection, the detector itself is usually the collecting aperture. As the particles interact with the detector material, ionization occurs, leading to light emission or charge release. Detection of the emitted light or generated current gives a measurement of the incident energy flux.

5- Generation of Electromagnetic Radiation

Electromagnetic radiation is generated by transformation of energy from other forms such as kinetic, chemical, thermal, electrical, magnetic, or nuclear. A variety of transformation mechanisms lead to electromagnetic waves over different regions of the electromagnetic spectrum. In general, the more organized (as opposed to random) the transformation mechanism is, the more coherent (or narrower in spectral bandwidth) is the generated radiation.

Radio frequency waves are usually generated by periodic currents of electric charges in wires, electron beams, or antenna surfaces. If two short, straight metallic wire segments are connected to the terminals of an alternating current generator, electric charges are moved back and forth between them. This leads to the generation of a variable electric and magnetic field near the wires and to the radiation of an electromagnetic wave at the frequency of the alternating current. This simple radiator is called a dipole antenna.

At microwave wavelengths, electromagnetic waves are generated using electron tubes that use the motion of high-speed electrons in specially designed structures to generate a variable electric/magnetic field, which is then guided by waveguides to a radiating structure. At these wavelengths, electromagnetic energy can also be generated by molecular excitation, as is the case in masers. Molecules have different levels of rotational energy. If a molecule is excited by some means from one level to a higher one, it could drop back to the lower level by radiating the excess energy as an electromagnetic wave.

Higher-frequency waves in the infrared and the visible spectra are generated by molecular excitation (vibrational or orbital) followed by decay. The emitted frequency is exactly related to the energy difference between the two energy levels of the molecules. The excitation of the molecules can be achieved by a variety of mechanisms such as electric discharges, chemical reactions, or photonic illumination.

Molecules in the gaseous state tend to have well-defined, narrow emission lines. In the solid phase, the close packing of atoms or molecules distorts their electron orbits, leading to a large number of different characteristic frequencies. In the case of liquids, the situation is compounded by the random motion of the molecules relative to each other.

Lasers use the excitation of molecules and atoms and the selective decay between energy levels to generate narrow-bandwidth electromagnetic radiation over a wide range of the electromagnetic spectrum ranging from UV to the high submillimeter.

Heat energy is the kinetic energy of random motion of the particles of matter. The random motion results in excitation (electronic, vibrational, or rotational) due to collisions, followed by random emission of electromagnetic waves during decay. Because of its random nature, this type of energy transformation leads to emission over a wide spectral band. If an ideal source (called a blackbody) transforms heat energy into radiant energy with the maximum rate permitted by thermodynamic laws, then the spectral emittance is given by Planck's formula as

$$S(\lambda) = \frac{2\pi hc^2}{\lambda^5}\frac{1}{e^{ch/\lambda kT} - 1} \qquad (27)$$

where h is Planck's constant, k is the Boltzmann constant, c is the speed of light, λ is the wavelength, and T is the absolute temperature in degrees Kelvin. Figure 2-12 shows the spectral emittance of a number of blackbodies with temperatures ranging from 2000° (temperature of the Sun's surface) to 300°K (temperature of the Earth's surface). The spectral emittance is maximum at the wavelength given by

$$\lambda_{max} = \frac{a}{T} \tag{28}$$

where a = 2898 µm°K. The total emitted energy over the whole spectrum is given by the Stefan–Boltzmann law:

$$S = \sigma T^4 \tag{29}$$

where $\sigma = 5.669 \times 10^{-8}\ \mathrm{Wm^{-2}K^{-4}}$. Thermal emission is usually unpolarized and extends through the total spectrum, particularly at the low-frequency end. Natural bodies are also characterized by their spectral emissivity $\varepsilon\ (\lambda)$, which expresses the capability to emit radiation due to thermal energy conversion relative to a blackbody with the same temperature.

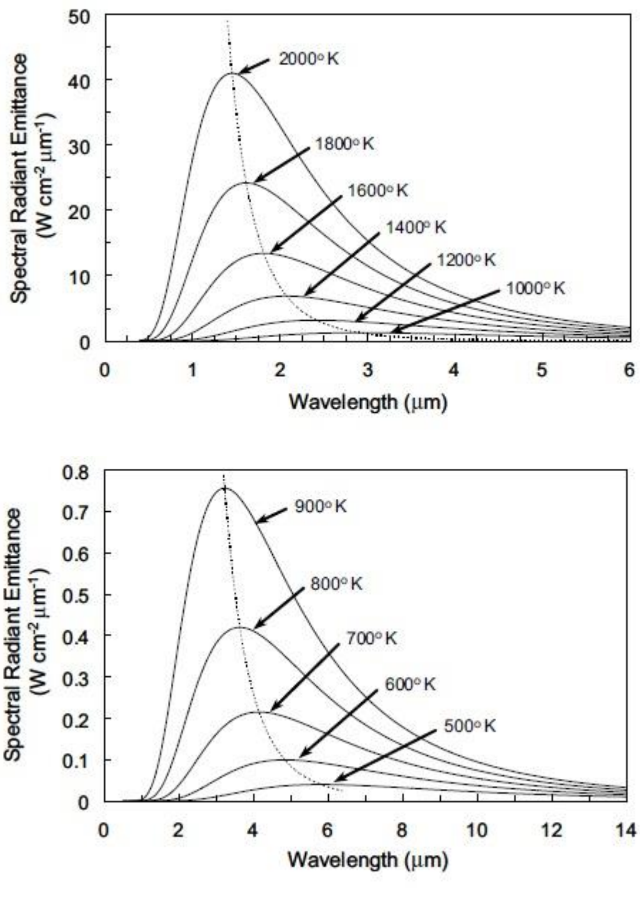


Figure (20). Spectral radiant emittance of a blackbody at various temperatures. Note the change of scale between the two graphs.

6-Interaction Mechanism throughout the Electromagnetic Spectrum

Starting from the highest spectral region used in remote sensing, gamma- and x-ray interactions with matter call into play atomic and electronic forces such as the photoelectric effect (absorption of photon with ejection of electron), Compton effect (absorption of photon with ejection of electron and radiation of lower-energy photon), and pair production effect (absorption of photon and generation of an electron–positron pair). The photon energy in this spectral region is larger than 40 eV (Fig. 21). This spectral region is used mainly to sense the presence of radioactive materials.

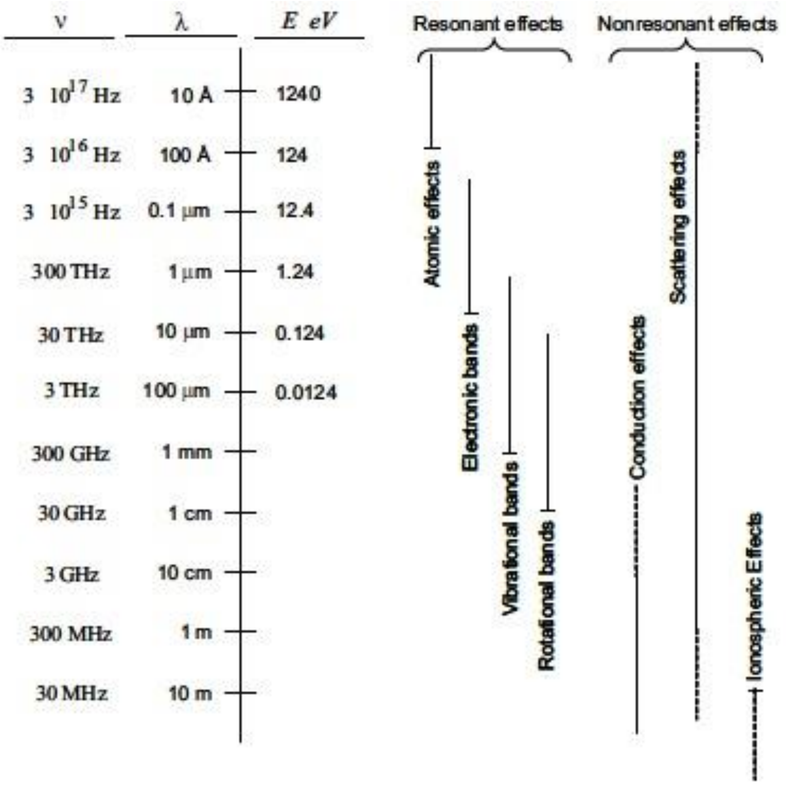


Figure (21). Correspondence of spectral bands, photon energy, and range of different wave–matter interaction mechanisms of importance in remote sensing. The photon energy in electron volts is given by E(eV) = 1.24/ λ, where λ is in μm.

In the ultraviolet region (photon energy between 3 eV and 40 eV), the interactions call into play electronic excitation and transfer mechanisms, with their associated spectral bands. This spectral region is used mostly for remote sensing of the composition of the upper layers of the Earth and planetary atmospheres. An ultraviolet spectrometer was flown on the Voyager spacecraft to determine the composition and structure of the upper atmospheres of Jupiter, Saturn, and Uranus.

In the visible and near infrared (energy between 0.2 eV and 3 eV), vibrational and electronic energy transitions play the key role. In the case of gases, these interactions usually occur at well-defined spectral lines, which are broadened due to the gas pressure and temperature. In the case of solids, the closeness of the atoms in the crystalline structure leads to a wide variety of energy transfer phenomena with broad interaction bands. These include molecular vibration, ionic vibration, crystal field effects, charge transfer, and electronic conduction. Table (2) shows the interaction mechanisms between matter and wave in the electromagnetic spectrum.

Table (2): Wave–Matter Interaction Mechanisms across the Electromagnetic Spectrum

Spectral region	Main interaction mechanisms	Examples of remote sensing applications
Gamma rays, x-rays	Atomic processes	Mapping of radioactive materials
Ultraviolet	Electronic processes	Presence of H and He in atmospheres
Visible and near infrared	Electronic and vibration molecular processes	Surface chemical composition, vegetation cover, and biological properties
Mid-infrared	Vibrational, vibrational–rotational molecular processes	Surface chemical composition, atmospheric chemical composition
Thermal infrared	Thermal emission, vibrational and rotational processes	Surface heat capacity, surface temperature, atmospheric temperature, atmospheric and surface constituents
Microwave	Rotational processes, thermal emission, scattering, conduction	Atmospheric constituents, surface temperature, surface physical properties, atmospheric precipitation
Radio frequency	Scattering, conduction, ionospheric effect	Surface physical properties, subsurface sounding, ionospheric sounding

7- Basic Concepts of atmospheric remote sounding

The techniques of atmospheric sounding can be divided into three general categories: occultation, scattering, and emission.

In the case of occultation techniques, the approach is to measure the changes that the atmosphere imparts on a signal of known characteristics as this signal propagates through a portion of the atmosphere. The signal source can be the sun, a star, or a man-made source such as a radio or radar transmitter. The geometry corresponds usually to limb sounding. Figure 22-*a* shows the case in which the sun is used as a source, whereas Figure 22-*b* shows the case in which a spacecraft radio transmitter is the source. This latter case is commonly used in planetary occultation. A special configuration is shown in Figure 22-*c* in which the surface is used as a mirror to allow the signal to pass through the total atmosphere twice.

In the case of scattering, the approach is to measure the characteristics of the scattered waves in a direction or directions away from the incident wave direction. The source can be the sun, as in Figure 8-13*d*, or man-made, as in Figure 22-*e*.

In the case of emission, the radiation source is the atmosphere itself (Figs. 22-*f* and *g*), and the sensor measures the spectral characteristics and intensity of the emitted radiation. The different atmospheric sounding techniques aim at measuring the spatial and temporal variations of the atmospheric properties, specifically temperature profile, constituents nature and concentration, pressure, wind, and density profile.

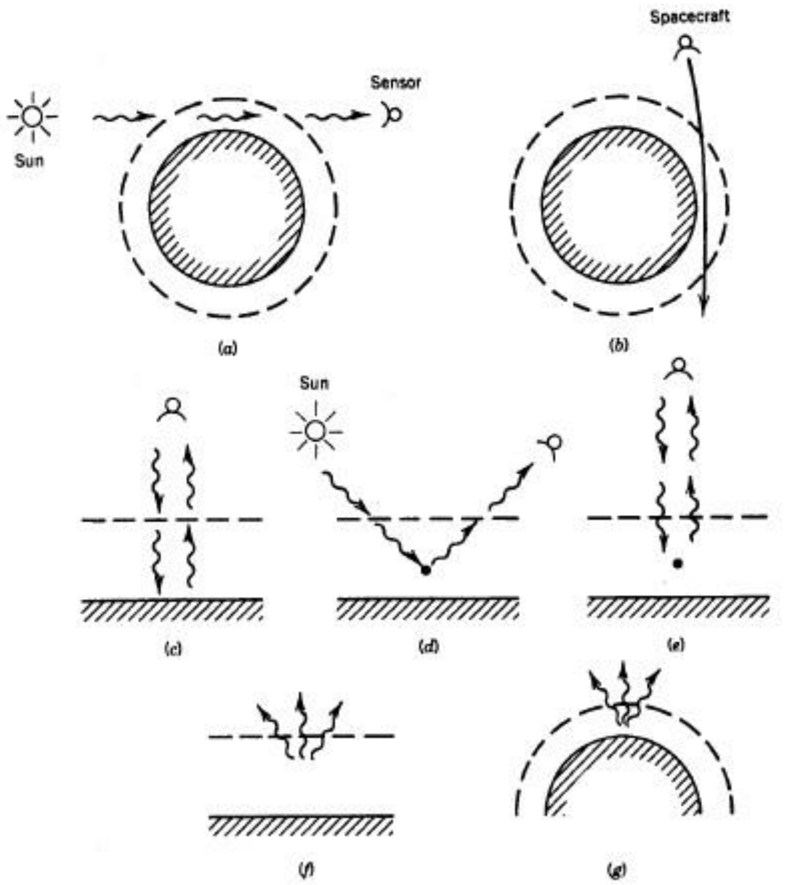


Figure (22): Different configurations for atmospheric sounding (see text for explanation).

7-1. Basic concept of Temperature Sounding

If a sensor measures the radiation emitted from gases whose distribution is well known, such as carbon dioxide or molecular oxygen in the Earth's atmosphere, then the radiance can be used to derive the temperature :

$$\psi_t(z) = \alpha_a(z)\, B\big(\nu, T(z)\big) \qquad (30)$$

Where ψ_t: thermal source term

α_a: extinction the absorption coefficient

B: Blank function (see equation 27)

At first glance, it seems that only the mean temperature can be derived because the radiation detected at any instant of time is a composite of waves emitted from all the different layers in the atmosphere. However, if we can measure the radiance variation as a function of frequency near a spectral line, the temperature vertical profile can be derived. This can be explained as follows.

The contribution from the layers at the top of the atmosphere is very small because the density (i.e., number of radiating molecules) is low. As we go deeper in the atmosphere, the contribution increases because of the higher atmospheric density. However, for the deep layers near the surface, even though the radiation source is the largest, the emitted radiation has to traverse the whole atmosphere, where it gets absorbed. Thus, its net contribution to the total radiance is small. This implies that, for a certain atmospheric optical thickness, there is an optimum altitude layer for which the combination of gas density (i.e., source strength) and attenuation above it are such that this layer contributes most to the total radiance. If the optical thickness changes, then the altitude of the peak contribution changes. Thus, if we observe the radiance at a number of neighboring frequencies for which the optical thickness varies over a wide range (this occurs when we look around an absorption spectral band), the altitude of the contribution peak will vary, thus allowing temperature measurement at different altitudes.
In order to get an accurate temperature profile, the absorption band used should have the following properties:

1. The emitting constituent should have a known mixing ratio and preferably be uniformly mixed in the atmosphere. This is the case for molecular oxygen and carbon dioxide in the Earth's atmosphere up to 100 km. The most commonly used bands are the 60 GHz band for oxygen and the 15 µm and 3.4 µm infrared bands for carbon dioxide. In the case of the Martian and Venusian atmospheres, CO2 infrared bands can be used. In the case of Jupiter, methane is uniformly mixed and its 7.7 µm infrared line can be used.
2. The absorption band involved should not be overlapped by bands from other atmospheric constituents. The 60 GHz oxygen line in the Earth's atmosphere, the 15 µm CO2 line in the Earth, Mars, and Venus atmospheres, and the 7.7 µm CH4 line in the Jovian atmosphere satisfy this requirement.
3. Local thermodynamic equilibrium should apply so that the Planck emission law is appropriate. This is usually the case in the lower 80 km of the Earth's atmosphere.
4. The wavelength should be long enough such that the scattered solar radiation is insignificant compared to the thermal emission. This is always the case in the microwave, millimeter, and thermal infrared part of the spectrum.

7-2. Basic concept for Composition Sounding

The identification of atmospheric constituents is usually based on detecting the presence of a spectral line or lines associated with a certain molecule. The spectral signature is in effect the "fingerprint" of a gaseous constituent.
In order to determine the abundance of a constituent, a more detailed analysis of the spectral signature is required. The line strength is usually related to the number density of molecules. This usually requires knowledge of the local pressure and temperature. Once the temperature is derived using the radiance from a homogeneously mixed constituent as discussed earlier, the corresponding abundance profiles can be derived by measuring the spectral radiance around other spectral lines. The "sounding" spectral lines should satisfy

the same properties discussed in the previous section except for the first one, which is relevant only to temperature sounding.

7-3. Basic concept of Pressure Sounding

The total columnar absorption is strongly related to the columnar mass of a constituent in the atmosphere, particularly near a resonant line of the constituent. If the constituent is homogeneously mixed in the atmosphere, its total mass is then directly proportional to the surface pressure. Thus, surface pressure sounding can be achieved by devising a technique to measure the total columnar absorption of a homogeneously mixed gas such as oxygen in the Earth's atmosphere.

7-3. Basic concept of Density Sounding.

The atmospheric refractivity N is directly proportional to the atmospheric density. Thus, one approach is to derive the refractivity profile as a function of altitude. This is done to derive the density profile of planetary atmospheres using the refraction of the radio communication signal as orbiting or fly by spacecraft are occulted by the atmosphere.

7-4. Basic concept of Wind Measurement.

The simplest technique for wind measurement is to take a time series of cloud hotographs, which would allow derivation of the wind field at cloud level. In order to get the wind field at any other altitude, the Doppler effect is used. The Multi-angle Imaging SpectroRadiometer (MISR) instrument that flies on the Terra spacecraft uses a series of images taken forward and aft of the spacecraft at different look angles and image matching algorithms to measure vector winds at the cloud levels.

Any molecule in motion will have its spectral line shifted by the Doppler effect. The Doppler shift is equal to

$$\Delta \nu = \frac{V}{\lambda} cos\theta \qquad (31)$$

where λ is the line wavelength and $V \cos \theta$ is the molecule velocity along the line of observation. Thus, by accurately measuring the line center for a known atmospheric constituent and comparing it to the frequency of the line for the same constituent in a static case, one of the velocity components can be derived. To illustrate, if a carbon dioxide molecule is moving at a line of sight velocity of 1 m/sec, the 15 µm line will have a frequency shift of

$$\Delta \upsilon = \frac{1}{15 \times 10^{-6}} \cong 66.7\ KHz$$

which is small but measurable. If we use the oxygen line at 60 GHz, then

$$\Delta\upsilon = \frac{1}{5\times10^{-3}} = 200\ KHz$$

Another technique for wind measurement, also based on the Doppler effect, uses the scattered wave from an illuminating laser or radar beam. The incident wave is scattered by

moving particles and the returned signal is shifted by a frequency $\Delta\upsilon$ given by

$$\Delta\nu = \frac{2V}{\lambda} cos\theta \qquad (32)$$

The returned signal is then mixed with a reference identical to the transmitted signal to derive the Doppler shift.

8-The History of Radar

Radar term is the abbreviation of RAdio Detecting And Ranging, i.e. finding and positioning a target and determining the distance between the target and the source by using radio frequency. This term was first used by the U.S. Navy in 1940 and adopted universally in 1943. It was originally called Radio Direction Finding (R.D.F.) in England.

The history of radar includes the various practical and theoretical discoveries of the 18th, 19th and early 20th centuries that paved the way for the use of radio as not occur until World War II, the basic principle of radar detection is almost as old as the means of communication. Although the development of radar as a stand-alone technology did subject of electromagnetism itself. Some of the major milestones of radar history are as follows:

· **1842** It was described by Christian Andreas Doppler that the sound waves from

a source coming closer to a standing person have a higher frequency while the sound waves from a source going away from a standing person have a lower frequency. That approach is valid for radio waves, too. In other words, observed frequency of light and sound waves was affected by the relative motion of the source and the detector. This phenomenon became known as the Doppler Effect.

• 1860 Electric and magnetic fields were discovered by Michael Faraday

• 1864Mathematical equations of electromagnetism were determined by Jame

Clark Maxwell. Maxwell set forth the theory of light must be accepted as a

electromagnetic wave. Electromagnetic field and wave were put forth

consideration by Maxwell.

. 1925 The first application of the pulse technique was used to measure distanc

by G. Breit and M. Truve.

· **1940** Microwaves were started to be used for long-range detection.

· **1947** The first weather radar was installed in Washington D.C. on February 14.

· **1950** Radars were put into operation for the detection and tracking of weather phenomena such as thunderstorms and cyclones.

· **1990.s** A dramatic upgrade to radars came in with the Doppler radar.

8-1 Basic Radar Terms

It seems beneficial to give at least the definition of some basic radar terms to be able to understand the theory and operation of radars. The common definitions of basic terms which will be talked about frequently during that training course are given below:

a) Frequency (f)

Frequency refers to the number of completed wave cycles per second

b) Phase (δ)

Phase of an electromagnetic wave is essentially the fraction of a full wavelength a particular point is from some reference point measured in radians or degrees.

c) Bandwidth (BW)

Bandwidth is the frequency difference between the upper and lower frequencies of electromagnetic radiation. It is expressed in units of Hertz (Hz).

d) Wavelength (λ)

This is distance from wavecrest to wavecrest (or trough to trough) along an electromagnetic wave.s direction of travel is called wavelength. Unit of wavelength is generally centimetre.

e) Pulse width (τ)

Pulse width is the time interval between the leading edge and trailing edge of a pulse at a point where the amplitude is 50% of the peak value. It is expressed in units of microseconds.

f) PRF and PRT

Pulse repetition frequency is the number of peak power pulses transmitted per second. Pulse repetition time is the time interval between two peak pulses.

g) Duty Factor/Duty Cycle

Duty cycle is the amount of time radar transmits compare to its listening to receiving time. The ratio is sometimes expressed in per cent. It can be determined by multiplying PRF and Pulse width or, by dividing the Pulse width with PRT**.** It has not any units

h) Beam width (θ)

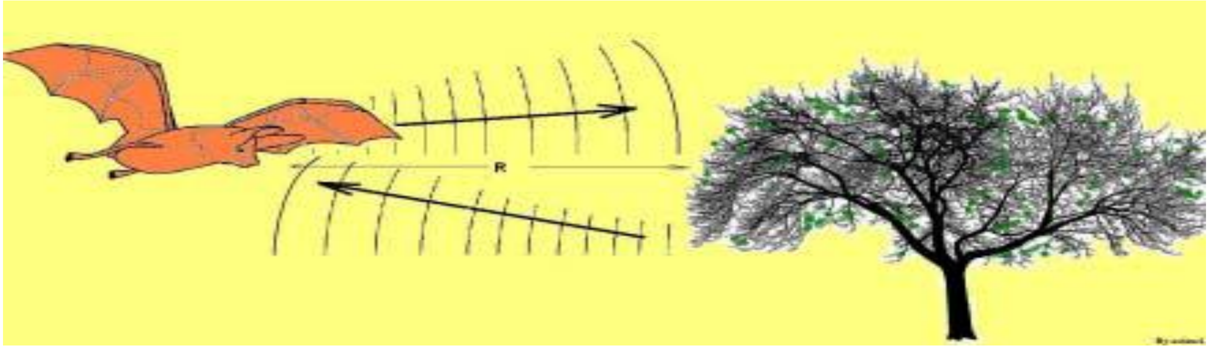
It is defined as the angle between the half-power (3 dB) points of the main lobe, when referenced to the peak effective radiated power of the main lobe. Unit is degree .

Table (3): Some Nomenclature

Name	Symbol	Units	Typical values
Transmitted frequency	f_t	MHz, GHz	1000-12500 MHz
wavelength	λ	cm	3-10 cm
Pulse duration	T	µsec	1 µsec
Pulse length	H	m	150-300 m ($h=c.\tau$)
Pulse repetition frequency	PRF	sec^{-1}	1000 sec^{-1}
interpulse period	T	millisecond	1 millisecond
peak transmitted power	P_t	MW	1 MW
average power	P_{avg}	kW	1 kW ($P_{avg} = P_t\, \tau$ PRF)
received power	P_r	mW	10^{-6} mW

8-3 Operation Principle of Radar

Operation principle of radar is very simple in theory and very similar to the way which bats use naturally to find their path during their flight **(Figure 2)**. Bats use a type of radar system by emitting ultrasonic sounds in a certain frequency (120 KHz) and hearing the echoes of these sounds. These echoes make them enable to locate and avoid the objects in their path.



In the radar systems, an electromagnetic wave generated by the transmitter unit is transmitted by means of an antenna and the reflected wave from the objects (echo) is received by the same antenna and after processing of the returned signal, a visual indication is displayed on indicators. After a radio signal is generated and emitted by a combination of a transmitter and an antenna, the radio waves travel out in a certain direction in a manner similar to light or sound waves. If the signals strike an object, the waves are reflected and the reflected waves travel in all directions depending of the surface of the reflector. The term reflectivity refers to the amount of energy returned from an object and is dependent on the size, shape and composition of the object. A small portion of the reflected waves return to the location of the transmitter originating them where they are picked up by the receiver antenna. This signal is amplified and displayed

on the screen of the indicators, e.g. PPI (Plan Position Indicator).This simple approach can be achieved by means of many complex process including hardware and software components.

8-4 Radar Equation

The fundamental relation between the characteristics of the radar, the target and the received signal is called the radar equation and the theory of radar is developed based on that equation.

$$\mathrm{P_r} = \frac{P_t G^2 \theta^2 H \Pi^3 K^2 L}{1024(\ln 2)\lambda^2} x \frac{Z}{R^2} \qquad (33)$$

This equation involves variables that are either known or are directly measured. There is only one value, **Pr,** that is missing but it can be solved for mathematically. Below is the list of variables, what they are and how they are measured.

Pr: Average power returned to the radar from a target. The radar sends pulses and then

measures the average power that is received in those returns. The radar uses multiple pulses since the power returned by a meteorological target varies from pulse to pulse. This is an unknown value of the radar but it is one that is directly calculated.

Pt: Peak power transmitted by the radar. This is a known value of the radar. It is important to know because the average power returned is directly related to the transmitted power.

G: Antenna gain of the radar. This is a known value of the radar. This is a measure of the

antenna's ability to focus outgoing energy into the beam. The power received from a given

target is directly related to the square of the antenna gain.

θ : Angular beam width of radar. This is a known value of the radar. Through the Robert-Jones equation it can be learned that the return power is directly related to the square of the angular beam width. The problem becomes that the assumption of the equation is that precipitation fills the beam for radars with beams wider than two degrees. It is also an invalid assumption for any weather radar at long distances. The lower resolution at great distances is called the aspect ratio problem.

H or (h): Pulse Length of the radar. This is a known value of the radar. The power received from a meteorological target is directly related to the pulse length.

K: This is a physical constant. This is a known value of the radar. This constant relies on the dielectric constant of water. This is an assumption that has to be made but also can cause some problems. The dielectric constant of water is near one, meaning it has a good reflectivity. The problem occurs when you have meteorological targets that do not share that reflectivity. Some examples of this are snow and dry hail since their constants are around 0.2.

L: This is the loss factor of the radar. This is a value that is calculated to compensate for attenuation by precipitation, atmospheric gases and receiver detection limitations. The attenuation by precipitation is a function of precipitation intensity and wavelength. For atmospheric gases, it is a function of elevation angle, range and wavelength. Since all of these accounts for a 2dB loss, all signals are strengthened by 2 dB.

λ: This is the wavelength of the transmitted energy. This is a known value of the radar. The amount of power returned from a precipitation target is inversely since the short wavelengths are subject to significant attenuation. The longer the wavelength, the less attenuation caused by precipitation.

Z: This is the reflectivity factor of the precipitate. This is the value that is solved for mathematically by the radar. The number of drops and the size of the drops affect this value. This value can cause problems because the radar cannot determine the size of the precipitate.The size is important since the reflectivity factor of a precipitation target is determined by raising each drop diameter in the sample volume to the sixth power and then summing all those values together. A ¼" drop reflects the same amount of energy as 64 1/8" drops even though there is 729 times more liquid in the 1/8" drops.

R: This is the target range of the precipitate. This value can be calculated by measuring the time it takes the signal to return. The range is important since the average power return from a target is inversely related to the square of its range from the radar. The radar has to normalize the power returned to compensate for the range attenuation.

Using a relationship between Z and R, an estimate of rainfall can be achieved. A base equation that can be used to do this is Z=200*R1.6. This equation can be modified at the user's request to a better fitting equation for the day or the area.

8-5 How to derive radar equation?

It may be interesting for somebody so, a general derivation steps of radar equation is given below. Our starting point will be flux calculations.

Flux Calculations - Isotropic Transmit Antenna

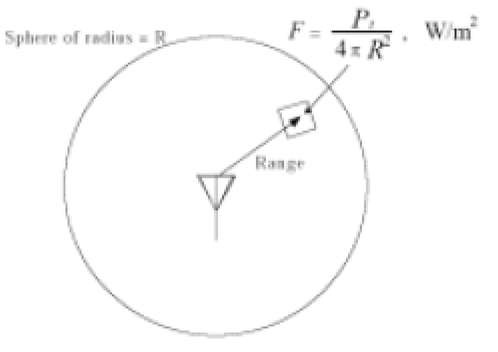


Figure (34): Flux at Distance R

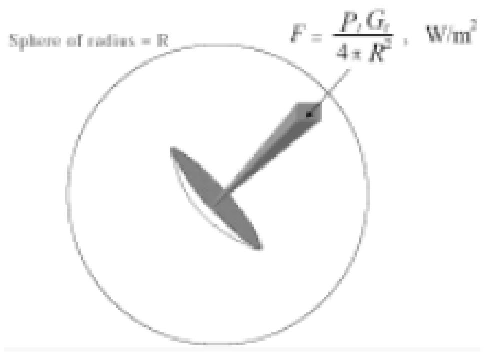
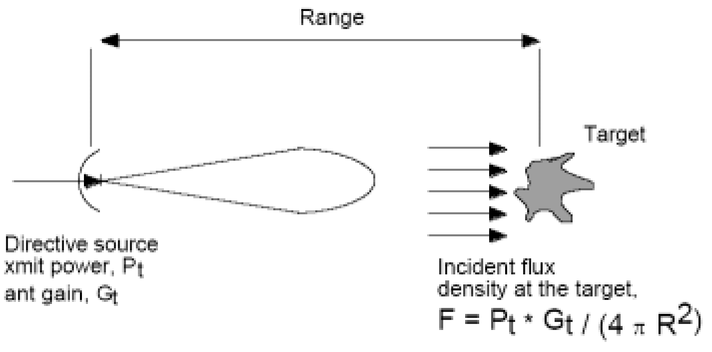


Figure (35): Flux at Distance R with Gain.

Radar Signal at Target, Incident power flux density from a Directive Source



Figure(36): Incident Power Flux Density from a Directive Source.

Echo Signal at Target, Backscattered power from the target

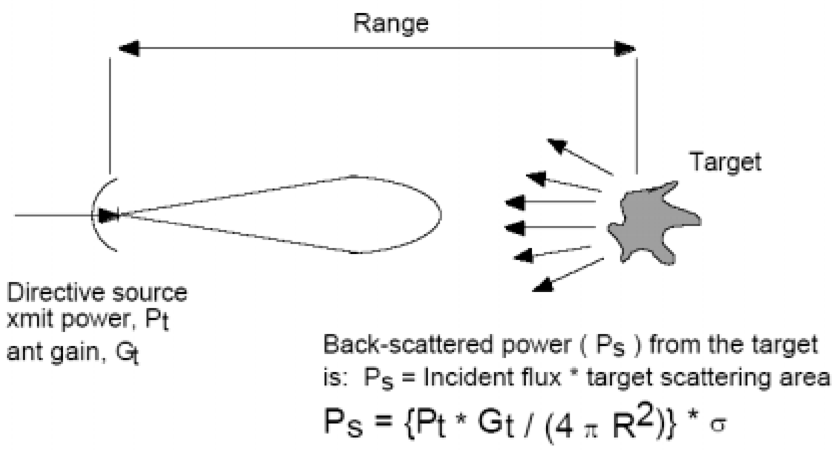


Figure (37): Power Back Scattered from Target with Cross Section.

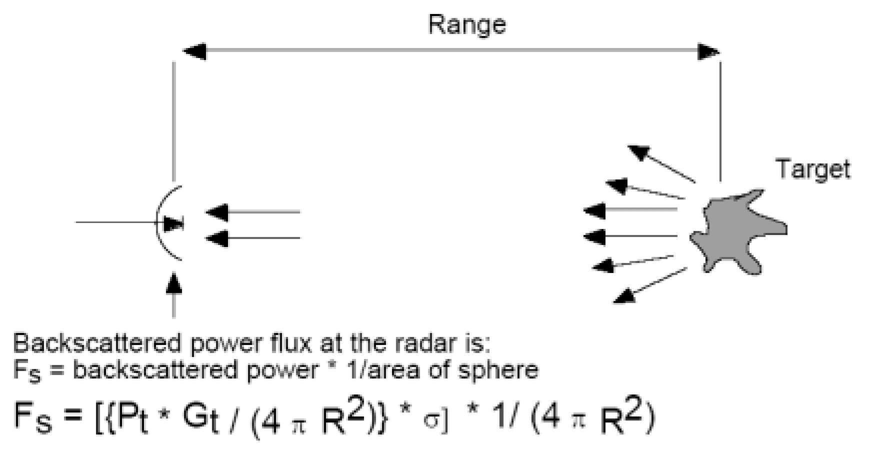


Figure (38): Flux Back Scattered from Target at Radar.

Radar Cross Section

The radar cross section (σ) of a target is the .equivalent area. of a flat-plate mirror:

♦ That is aligned perpendicular to the propagation direction (i.e., reflects the signal directly back to the transmitter) and

♦ That results in the same backscattered power as produced by the target

Radar cross section is extremely difficult to predict and is usually measured using scaled models of targets.

Target Echo Signal at Radar Received (echo) power at the radar

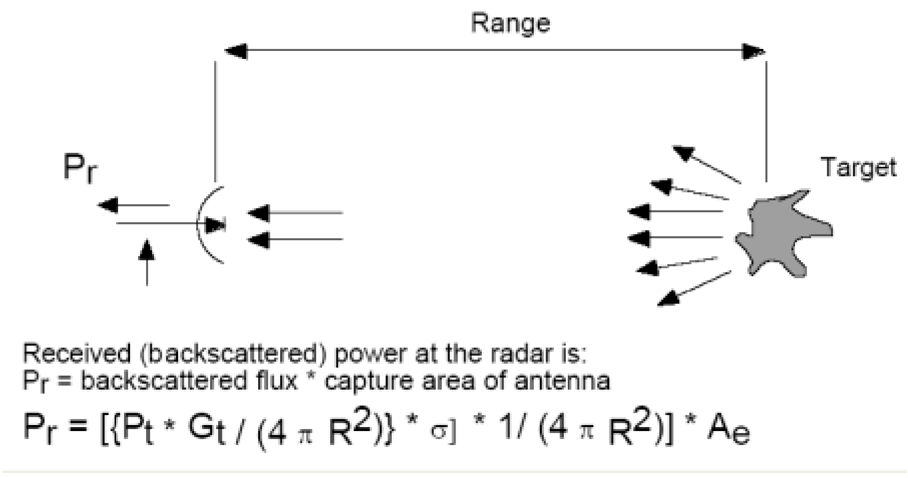


Figure (39):Received Power at Radar.

Relationship between Antenna Aperture and Gain

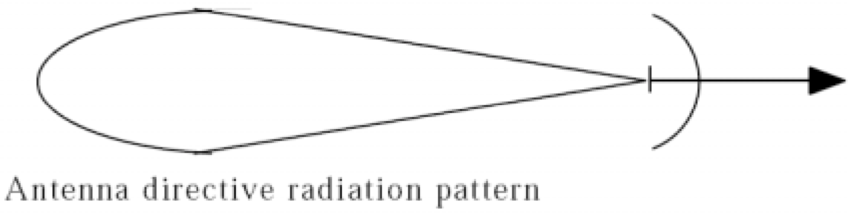


Figure (40): Antenna Aperture and Gain.

$$G_r = \frac{4\pi A_e}{\lambda^2}, \text{ power ratio}$$

$$A_e = \rho_a * A, \; meters^2$$

Where A= the physical aperture area of the antenna

ρ_a = the aperture collection efficiency

λ = wave length electromagnetic = c/freq

Idealized Radar Equation - no system losses

$$P_r = P_t * G_t * G_r \left[\frac{\lambda^2}{(4\pi)^3 R^4} \right] * \sigma \ , \ watts$$

Since the antenna gain is the same for transmit and receive, this becomes:

$$P_r = P_t * G^2 \left[\frac{\lambda^2}{(4\pi)^3 R^4} \right] * \sigma \ , \ watts$$

Practical Radar Equation - with system losses for point targets

$$P_r = P_t * G^2 \left[\frac{\lambda^2}{(4\pi)^3 R^4} \right] * \sigma * L_{sys} \ , \ watts$$

Where:

Lsys is the system losses expressed as a power ratio,
- P_r is the average received power,
- P_t is the transmitted power,
- G is the gain for the radar,
- λ is the radar's wavelength,
- σ is the targets scattering cross section,
- R is the range from the radar to the target.

The radar equation for a point target is simply given below:

$$\overline{P_r} = \frac{P_t G^2 \lambda^2 \sigma}{(4\pi)^3 R^4}$$

5-9 Radar Equation for Distributed Targets

Thus far, we've derived the radar equation for a point target. This is enough if you are interested in point targets such as airplanes. However, in a thunderstorm or some area of Precipitation, we do not have just one target (e.g., raindrop), we have many. Thus we need to derive the radar equation for distributed targets. So let review the Radar Pulse Volume.

Radar Pulse Volume

First, let's simplify the real beam according to the Figure (41):

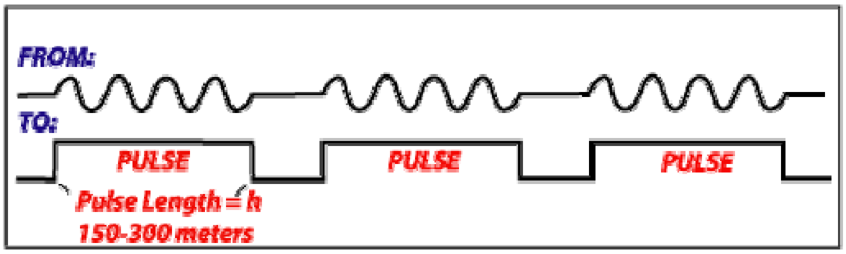


Figure (41): Radar Pulses.

What does a "three-dimensional" segment of the radar beam look like?

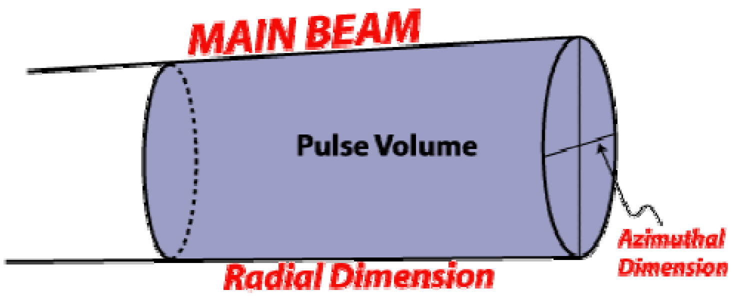


Figure (42): Radar Main Beam and Pulse Volume.

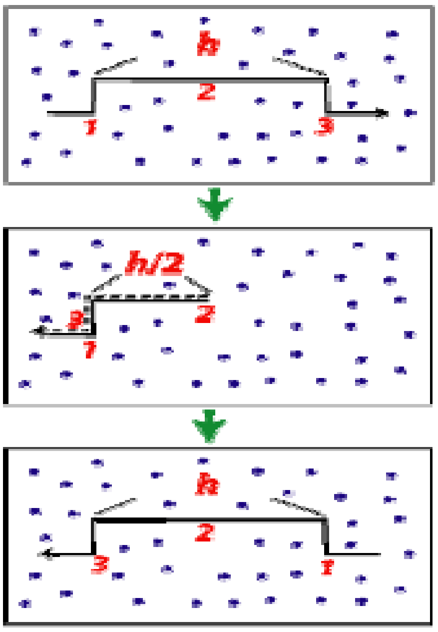


Figure (43): Form of Transmit and Received Signal.

So, the volume of the pulse volume is:

$$\Phi^{\pi}\frac{R\theta}{2}\frac{R\phi}{2}\frac{h}{2}$$

For a circular beam, then $\theta = \Phi$, the pulse volume becomes:

$$\pi\frac{R^2\theta^2 h}{8}$$

Before we derive the radar equation for the distributed targets situation, we need to make some assumptions:

1) The beam is filled with targets.

2) Multiple scattering is ignored

3) Total average power is equal to sum of powers scattered by individual particles.

Recall the radar equation for a single target:

$$\overline{P}_r = \frac{P_t G^2 \lambda^2 \sigma}{(4\pi)^3 R^4} \qquad (34)$$

For multiple targets, radar equation (34) can be written as:

$$\overline{P}_r = \frac{P_t G^2 \lambda^2}{(4\pi)^3}\sum_{1}\frac{\sigma_1}{R_i^4} \qquad (35)$$

where the sum is over all targets within the pulse volume.

If we assume that *h/2 << ri*,

Figure (44): Pulse Volume

Then (35) can be written as:

$$\overline{P}_r = \frac{P_t G^2 \lambda^2}{(4\pi)^3 R_i^4} \sum_i \sigma_i \qquad (36)$$

It is advantageous to sum the backscattering cross sections over a unit volume of the total pulse volume.

Hence the sum in (36) can be written as:

$$\sum_i \sigma_i = \left(\sum_i \frac{\sigma_i}{unitvolume} \right) totalvolume$$

where the total volume is the volume of the pulse:

$$\sum_i \sigma_i = \left(\sum_i \frac{\sigma_i}{unitvolume} \right) \pi \frac{R^2 \theta^2 h}{8}$$

Note that:

Pr is proportional to R^{-2} for distributed targets.

Pr is proportional to R^{-4} for point targets.

5-10 Radar Reflectivity

The sum of all backscattering cross sections (per unit volume) is referred to as the radar reflectivity (**η**). In other words,

$$\sum_i \sigma_i = \eta \qquad (37)$$

In terms of the radar reflectivity, the radar equation for distributed targets can be written as

$$\overline{P}_r = \left(\frac{P_t G^2 \lambda^2 \theta^2 h}{512 \pi^2 R^2} \right) \eta \qquad (38)$$

All variables in (38), except η are either known or measured.
Now, we need to add a fudge factor due to the fact that the beam shape is Gaussia

Hence, (38) becomes;

$$\overline{P}_r = \frac{1}{2\ln 2}\left(\frac{P_t G^2 \lambda^2 \theta^2 h}{512\pi^2 R^2}\right)\eta \tag{39}$$

Complex Dielectric Factor

The backscattering cross section (σ) is a function of the object characteristics and the size of the illuminated area. It can be written as:

$$\sigma = \frac{\pi^5 |K|^2 D^6}{\lambda^4} \tag{40}$$

Where:

· *D* is the diameter of the target,

· λ is the wavelength of the radar,

· *K* is the complex dielectric factor,

σ is some indication of how good a material is at backscattering radiation

For water $|K|^2$ =0.93

For ice $|K|^2$ =0.197

Notice that the value for water is much larger than for ice. All other factors the same; this creates a 5dB difference in returned power.

So, let's incorporate this information into the radar equation.

$$\text{As:}\ \eta = \sum_i \sigma_i = \sum_i \frac{\pi^5 |K|^2 D_i^6}{\lambda^4} \tag{41}$$

Taking the constants out of the sum;

$$\eta = \frac{\pi^5 |K|^2}{\lambda^4} \sum_i D_i^6 \tag{42}$$

Remember that the sum is for a unit volume. Substituting (41) into (38) gives:

$$\overline{P}_r = \frac{1}{2\ln 2}\left(\frac{P_t G^2 \lambda^2 \theta^2 h}{512\pi^2 R^2}\right)\frac{\pi^5 |K|^2}{\lambda^4}\sum_i D_i^6 \tag{43}$$

Simplifying terms gives:

$$\overline{P}_r = \frac{P_t G^2 \theta^2 \pi^3 h |K|^2}{1024 \ln 2 R^2 \lambda^2}\sum_i D_i^6 \tag{44}$$

Note the D^6 dependence on the average received power.

Radar Reflectivity Factor

In Equation (44), all variables except the summation term, are either known or measured.

We will now define the radar reflectivity factor, Z as:

$$Z = \sum_i \frac{D_i^6}{unitvolume} \tag{45}$$

Substituting (45) into (44) gives the radar equation for **distributed** targets:

$$\overline{P}_r = \frac{P_t G^2 \theta^2 \pi^3 h |K|^2 Z}{1024 \ln 2 \lambda^2 R^2} \tag{46}$$

5-11 Block Diagram of Radar

Radar systems, like other complex electronics systems, are composed of several major subsystems and many individual circuits. Although modern radar systems are quite complicated, you can easily understand their operation by using a basic radar block diagram.

Figure(45) below shows us the basic radar block diagram.

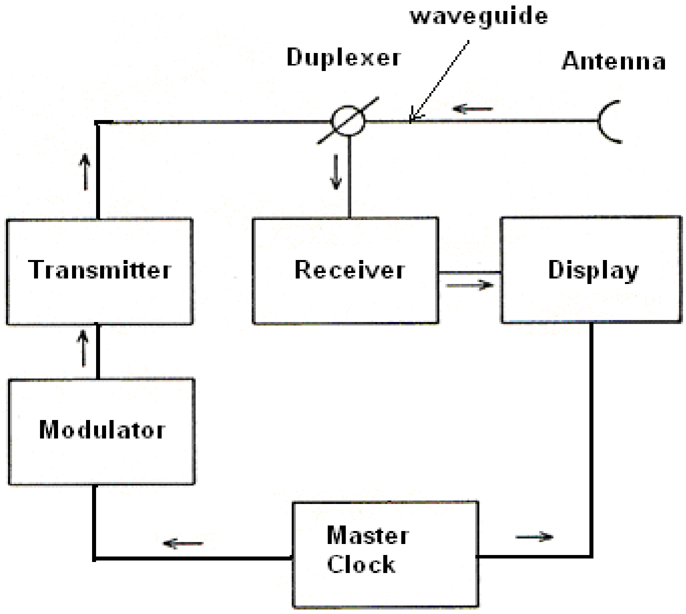


Figure (45): Basic Radar Diagram.

The parts of this block diagram in **Figure 14** are described below:

Master Clock/Computer: In older radars, this device was called the master clock. It would generate all of the appropriate signals and send them to the appropriate components of the radar. In modern radars, the function of the master clock has been taken over by the ubiquitous computer. Computers now control radars just as they control many other parts of modern technology.

Transmitter: The source of the EM radiation emitted by radar is the transmitter. It generates the high frequency signal which leaves the radar's antenna and goes out into the atmosphere. The transmitter generates powerful pulses of electromagnetic energy at precise intervals. The required power is obtained by using a high-power microwave oscillator (such as a magnetron) or a microwave amplifier (such as a klystron) that is supplied by a low- power RF source.

Modulator: The purpose of modulator is to switch the transmitter **ON** and **OFF** and to provide the correct waveform for the transmitted pulse. That is, the modulator tells the transmitter when to transmit and for what duration.

Waveguide: Figure (45) shows that the connecting the transmitter and the antenna is waveguide. This is usually a hollow, rectangular, metal conductor whose interior dimensions depend upon the wavelength of the signals being carried. Waveguide is put together much like the copper plumbing in a house. Long piece of waveguide are connected together by special joints to connect the transmitter/receiver and the antenna.

Antenna: The antennas are the device which sends the radar's signal into atmosphere. Most antennas used with radars are directional; that is, they focus the energy into a particular direction and not other directions. An antenna that sends radiation equally in all directions is called **isotropic antenna.**

Receiver: The receiver is designed to detect and amplify the very weak signals received by antenna. Radar receivers must be of very high quality because the signals that are detected are often very weak.

Display: There are many ways to display radar data. The earliest and easiest display for radar data was to put it onto a simple oscilloscope. After that A-scope was found. PPI and RHI are new techniques for displaying the radar data.

Duplexer: Duplexer, somebody called Transmit/Receive switch, is a special switch added to the radar system to protect the receiver from high power of the transmitter.

Of course this is a briefly explanation about components of a radar. Later on this course and in the other modules of this training document, besides these parts, all components will be explained in detail.

5-12 Commonly used imaging radar bands

Similarly to optical remote sensing, radar sensors operate with different bands.

For better identification, a standard has been established that defines various wavelength ranges using letters to distinguish among the various bands. In the description of different radar missions you will recognize the different wavelengths used if you see the letters. The European ERS mission and the Canadian Radarsat, for example, use C-band radar. Just like multispectral bands, different radar bands provide information about different object characteristics.

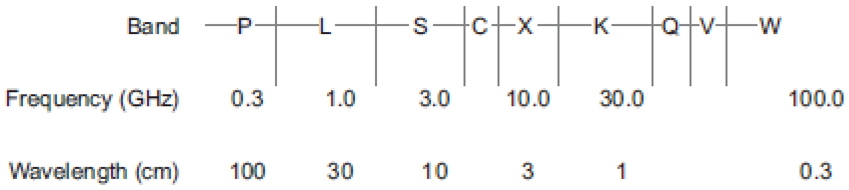


Figure (46): Microwave spectrum and band identification by letters first previous

5-13 Microwave polarizations

The polarization of an electromagnetic wave is important in the field of radar remote sensing. Depending on the orientation of the transmitted and received radar wave, polarization will result in different images (Figure 47). It is possible to work with horizontally, vertically or cross-polarized radar waves. Using different polarizations and wavelengths, you can collect information that is useful

for particular applications, for example, to classify agricultural fields. In radar system descriptions you will come across the following abbreviations:

- HH: horizontal transmission and horizontal reception,
- VV: vertical transmission and vertical reception,
- HV: horizontal transmission and vertical reception, and
- VH: vertical transmission and horizontal reception

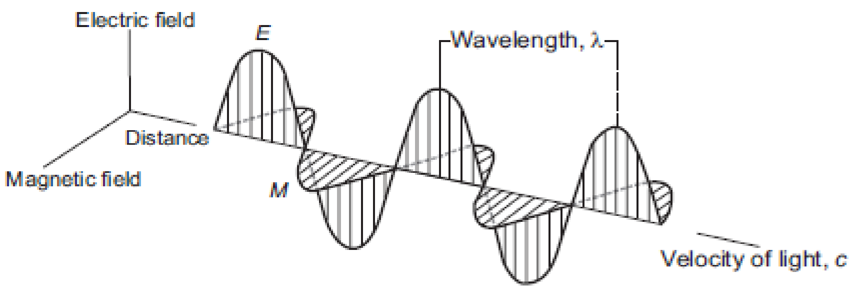


Figure 47: **A vertically polarized electromagnetic wave; the electric field's Variation occurs in the vertical plane in this example.**

5-14 RADAR SOUNDING OF RAIN

The use of radar sensors for rain detection, sounding, and tracking began in the early 1940s. As a radar pulse interacts with a rain region, echoes are returned that are proportional to the radar backscatter of the rain particles. As the pulse propagates deeper in the rain cell, additional echoes are returned at later times. Thus, a time analysis of the returned echo provides a profile of the rain intensity along the radar line of sight.

Three radar configurations are used to map a rain cell. The plan position indicator (PPI), the range height indicator (RHI), and the down looking profiler (see Fig.48). The PPI configuration corresponds to a horizontal cut around the sensor acquired by rotating the radar beam continuously, deriving an image of precipitation cells on a conical surface. The RHI gives a vertical cut in one plane acquired by scanning the beam up and down at a fixed azimuth angle. A combined PPI and RHI system would allow the acquisition of a three-dimensional image of the rain cells in the volume surrounding the sensor.

The down looking profiler corresponds to an airborne or spaceborne sensor, and it provides a vertical cut of the precipitating region along the flight line. Figure (49) gives an example of such data acquired at the X band with an airborne radar sensor. It clearly shows the top profile of the precipitation region and the bright band that corresponds to the melting level. If the radar beam of a down looking profiler is scanned back and forth across the track, then a wide volume of the atmosphere is mapped as the platform (aircraft or spacecraft) moves by.

If we consider the backscatter cross section of a volume of small particles, then [from Equation (42)]

$$\sigma = \frac{\pi^5}{\lambda^4}|K|^2 \sum D^6$$

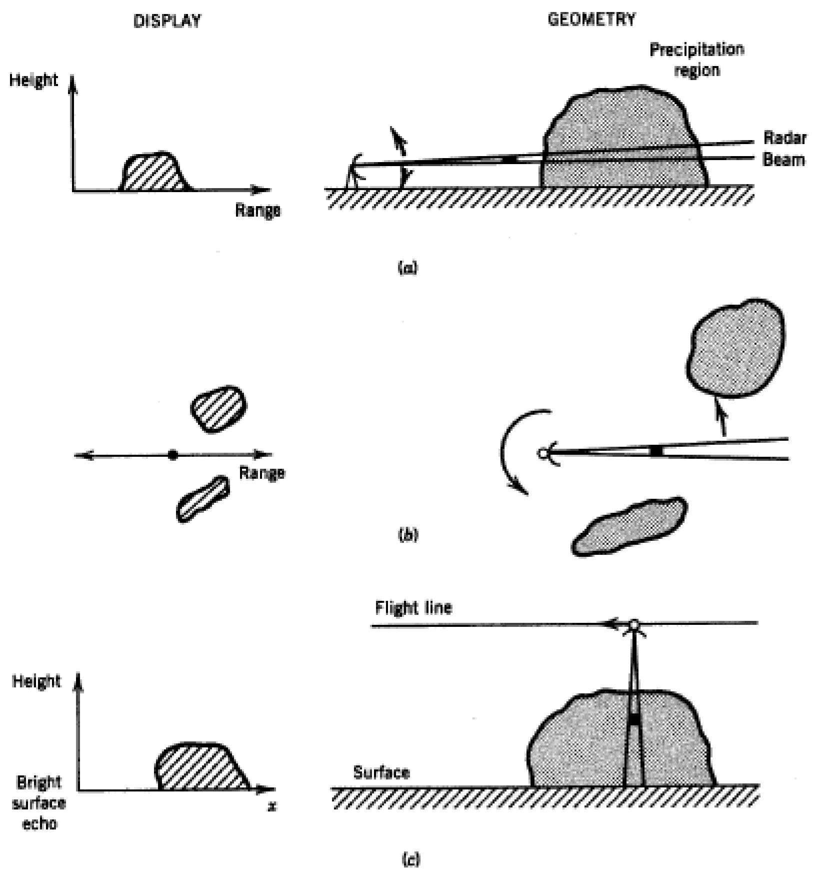


Figure (48): Three rain radar configurations. (*a*) Range height indicator, (*b*) plan position indicator,and (*c*) down looking profile.

where D = is the diameter of the particle. The summation term is commonly called the reflectivity factor Z:

$$Z = \sum D^6 \tag{47}$$

and is expressed in mm6/m3 or in m3. In the case of a distribution of particles,

$$Z = \int N(D) D^6 \, dD \tag{48}$$

where *N*(*D*) is the particles' size distribution. Numerous research activities have been conducted to relate the parameter *Z* to the rain rate *R* expressed in mm/hr and the liquid water content *M* expressed in mg/m3. The liquid water content is given by

Figure (49): Three rain radar configurations. (*a*) Range height indicator, (*b*) plan position indicator,and (*c*) downlooking profile.

$$M = \int_0^\infty \rho \frac{\pi D^3}{6} N(D)\, dD = \frac{\pi\rho}{6} \int_0^\infty D^3 N(D) d\, D \qquad (49)$$

where ρ is the water density. In the case of an exponential drop distribution,

$$N(D) = N_0 e^{-D/\overline{D}}$$

and, from Equations (48) and (49),

$$Z = 720 N_0 (\overline{D})^7$$

$$M = \pi\rho N_0 (\overline{D})^4$$

$$D^4 = \left(\frac{(M)}{(\pi\rho N0)}\right)$$

$$Z = \frac{720}{(\pi\rho)^{7/4} N_0^{3/4}} M^{7/4} \sim M^{1.75} \qquad (50)$$

If all the particles were of the same size, then

$$Z = ND^6$$

$$M = \frac{\pi\rho}{6} ND^3$$

$$\rightarrow Z = \frac{36}{(\pi\rho)^2 N} M^2 \sim M^2$$

Experimental investigations showed that for clouds, snow, and rain,

$$Z = CM^{\alpha}$$

where 1.8 < _ < 2.2, and *C* is a constant given by (when *M* is in mg/m3 and *Z* in mm6/m3):

$$\begin{array}{ll} C \simeq 5 \times 10^{-8} & \text{for clouds} \\ C \simeq 0.01 & \text{for snow} \\ C \simeq 0.08 & \text{for rain} \end{array}$$

If we assume that all the drops fall at the same velocity *V*, then the rain rate *R* is related to the liquid water content by

$$R = \frac{VM}{\rho}$$

Statistical measurements showed that $R \sim M^{0.9}$ and that

$$Z = BR^b$$

where 1.6 < *b* < 2 and *B* is a constant given by (when *R* is mm/hr):

B = 200 for rain

B = 2000 for snow

RADAR EQUATION FOR PRECIPITATION MEASUREMENT

From Equations (42) and (47), the radar backscatter of a unit volume of small particles is

$$\sigma = \frac{\pi^5}{\lambda^4} |K|^2 Z$$

If a radar sensor radiates a pulse of peak power P_t and length P_i through an antenna of area *A* and gain *G*, the power density *P*i at a distance *r* is

$$P_i = \frac{P_t G}{4\pi r^2} = \frac{P_t A}{\lambda^2 r^2}$$

The backscattered power *P*s at any instant of time corresponds to the return from the particles in a volume of depth $c\tau/2$ and area *S* equal to the beam footprint. Thus,

$$P = P_i \sigma \frac{Sc\tau}{2}$$

Where

$$S = \frac{\lambda^2 r^2}{A}$$

$$\rightarrow P_s = \frac{P_i \sigma \lambda^2 r^2 c\tau}{2A}$$

The received power *P*r is equal to the power collected by the antenna:

$$P_r = \frac{P_s}{4\pi r^2} A = \frac{P_i \lambda^2 c\tau \sigma}{8\pi}$$

$$\rightarrow P_r = \frac{\pi^4}{8} Z |K^2| \frac{P_t A c\tau}{\lambda^4 r^2}$$

In addition, *P*r should be reduced by the loss factor due to absorption and scattering along the path of length *r*.In the case of a coherent radar sensor, the received echo is mixed with the appropriate signal from the reference local oscillator in order to derive any shift in the echo’s frequency relative to the transmitted pulse. This corresponds to the Doppler shift resulting from the motion of the droplets relative to the sensor.

Satellite Characteristics: Orbits and Swaths

We learned in the previous section that remote sensing instruments can be placed on a variety of platforms to view and image targets. Although ground-based and aircraft platforms may be used, satellites provide a great deal of the remote sensing imagery commonly used today. Satellites have several unique characteristics which make them particularly useful for remote sensing of the Earth's surface. The path followed by a satellite is referred to as its **orbit**.

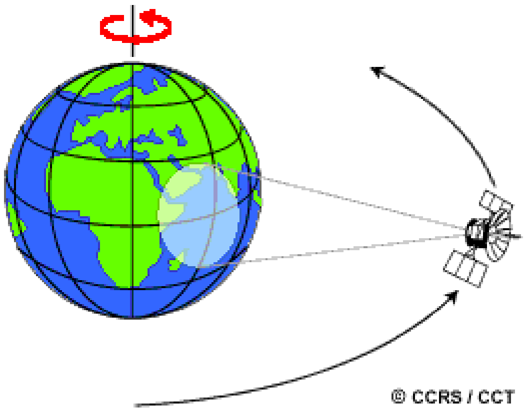


Figure (50): geostationary orbits

Satellite orbits are matched to the capability and objective of the sensor(s) they carry. Orbit selection can vary in terms of altitude (their height above the Earth's surface) and their orientation and rotation relative to the Earth. Satellites at very high altitudes, which view the same portion of the Earth's surface at all times have **geostationary orbits (**Figure 50). These geostationary satellites, at altitudes of approximately 36,000 kilometres, revolve at speeds which match the rotation of the Earth so they seem stationary, relative to the Earth's surface. This allows the satellites to observe and collect information continuously over specific areas. Weather and communications satellites commonly have these types of orbits. Due to their high altitude, some geostationary weather satellites can monitor weather and cloud patterns covering an entire hemisphere of the Earth.

Many remote sensing platforms are designed to follow an orbit (basically north-south) which, in conjunction with the earth's rotation (west-east), allows them to cover most of the earth's surface over a certain period of time. These are **nearpolar orbits**, so named for the inclination of the orbit relative to a line running between the North and South poles. Many of these satellite orbits are also **sun-synchronous** such that they cover each area of the world at a constant local time of day called **local sun time**. At any given latitude, the position of the sun in the sky as the satellite passes overhead will be the same within the same season. This ensures consistent illumination conditions when acquiring images in a specific season over successive years, or over a particular area over a series of days. This is an important factor for monitoring changes between images or for mosaicking adjacent images together, as they do not have to be corrected for different illumination conditions.

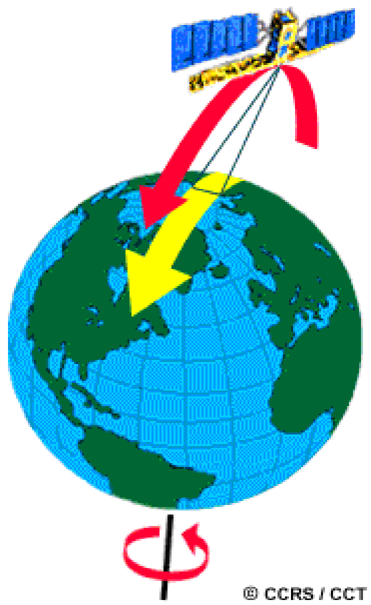
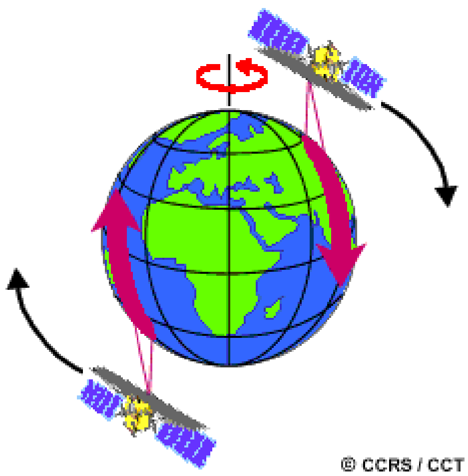
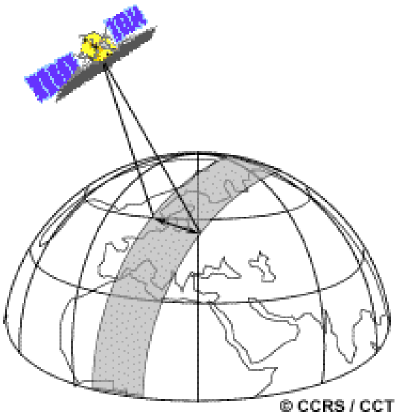


Figure (51): Nearpolar Orbits

Most of the remote sensing satellite platforms today are in near-polar orbits, which means that the satellite travels northwards on one side of the Earth and then toward the southern pole on the second half of its orbit. These are called **ascending and descending passes**, respectively. If the orbit is also sunsynchronous, the ascending pass is most likely on the shadowed side of the Earth while the descending pass is on the sunlit side. Sensors recording reflected solar energy only image the surface on a descending pass, when solar illumination is available. Active sensors which provide their own illumination or passive sensors that record emitted (e.g. thermal) radiation can also image the surface on ascending passes.



As a satellite revolves around the Earth, the sensor "sees" a certain portion of the Earth's surface. The area imaged on the surface, is referred to as the **swath**. Imaging swaths for space borne sensors generally vary between tens and hundreds of kilometers wide. As the satellite orbits the Earth from pole to pole, its east-west position wouldn't change if the Earth didn't rotate. However, as seen from the Earth, it seems that the satellite is shifting westward because the Earth is rotating (from west to east) beneath it. This apparent movement allows the satellite swath to cover a **new area with each consecutive pass**. The satellite's orbit and the rotation of the Earth work together to allow complete coverage of the Earth's surface, after it has completed one complete cycle of orbits.



If we start with any randomly selected pass in a satellite's orbit, an orbit cycle will be completed when the satellite retraces its path, passing over the same point on the earth's surface directly below the satellite (called the **nadir** point) for a second time. The exact length of time of the orbital cycle will vary with each satellite. The interval of time required for the satellite to complete its orbit cycle is not the same as the "**revisit period**". Using steerable sensors, an satellite-borne instrument can view an area (off-nadir) before and after the orbit passes over a target, thus making the 'revisit' time less than the orbit cycle time. The revisit period is an important consideration for a number of monitoring applications, especially when frequent imaging is required (for example, to monitor the spread of an oil spill, or the extent of flooding). In near-polar orbits, areas at high latitudes will be imaged more frequently than the equatorial zone due to the increasing **overlap in adjacent swaths** as the orbit paths come closer together near the poles (Figure (52)).

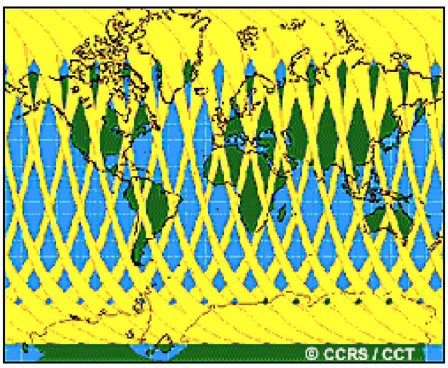


Figure (52): Orbit Paths

Weather Satellites/Sensors

Weather monitoring and forecasting was one of the first civilian (as opposed to military) applications of satellite remote sensing, dating back to the first true weather satellite, TIROS-1 (Television and Infrared Observation Satellite - 1), launched in 1960 by the United States. Several other weather satellites were launched over the next five years, in near-polar orbits, providing repetitive coverage of global weather patterns. In 1966, NASA (the U.S. National Aeronautics and Space Administration) launched the geostationary Applications Technology Satellite (ATS-1) which provided **hemispheric images** of the Earth's surface and cloud cover every half hour. For the first time, the development and movement of weather systems could be routinely monitored. Today, several countries operate weather, or meteorological satellites to monitor weather conditions around the globe. Generally speaking, these satellites use sensors which have fairly coarse spatial resolution (when compared to systems for observing land) and provide large areal coverage.

Their temporal resolutions are generally quite high, providing frequent observations of the Earth's surface, atmospheric moisture, and cloud cover, which allows for near-continuous monitoring of global weather conditions, and hence - forecasting. Here we review a few of the representative satellites/sensors used for meteorological applications.

GOES

The GOES (Geostationary Operational Environmental Satellite) System is the follow-up to the ATS series. They were designed by NASA for the National Oceanic and Atmospheric Administration (NOAA) to provide the United States National Weather Service with frequent, small-scale imaging of the Earth's surface and cloud cover. The GOES series of satellites have been used extensively by meteorologists for weather monitoring and

forecasting for over 20 years. These satellites are part of a global network of meteorological satellites spaced at approximately 70° longitude intervals around the Earth in order to provide near-global coverage. Two GOES satellites, placed in **geostationary orbits** 36000 km above the equator, each view approximately one-third of the Earth. One is situated at 75°W longitude and monitors North and South America and most of the Atlantic Ocean. The other is situated at 135°W longitude and monitors North America and the Pacific Ocean basin. Together they cover from 20°W to 165°E longitude. This GOES image covers a portion of the southeastern United States, and the adjacent ocean areas where many severe storms originate and develop. Image below (figure (53)) shows Hurricane Fran approaching the southeastern United States and the Bahamas in September of 1996.

Figure (53): Hurricane Fran

Two generations of GOES satellites have been launched, each measuring emitted and reflected radiation from which atmospheric temperature, winds, moisture, and cloud cover can be derived. The first generation of satellites consisted of GOES-1 (launched 1975) through GOES-7 (launched 1992). Due to their design, these satellites were capable of viewing the Earth only a small percentage of the time (approximately five per cent). The second generation of satellites began with GOES-8 (launched 1994) and has numerous technological improvements over the first series. They provide near-continuous observation of the Earth allowing more frequent imaging (as often as every 15 minutes). This increase in temporal resolution coupled with improvements in the spatial and radiometric resolution of the sensors provides timelier information and improved data quality for forecasting meteorological conditions.

GOES-8 and the other second generation GOES satellites have separate **imaging** and **sounding** instruments. The **imager** has five channels sensing visible and infrared reflected and emitted solar radiation. The infrared capability allows for day and night imaging. Sensor pointing and scan selection capability enable imaging of an entire hemisphere, or small-scale imaging of selected areas. The latter allows meteorologists to monitor specific weather trouble spots to assist in improved short-term forecasting. The imager data are 10-bit radiometric resolution, and can be transmitted directly to local user terminals on the Earth's surface. The accompanying table describes the individual bands, their spatial resolution, and their meteorological applications.

GOES Bands

Band	Wavelength Range (>µm)	Spatial Resolution	Application
1	0.52 - 0.72 (visible)	1 km	cloud, pollution, and haze detection; severe storm identification
2	3.78 - 4.03 (shortwave IR)	4 km	identification of fog at night; discriminating water clouds and snow or ice clouds during daytime; detecting fires and volcanoes; night time determination of sea surface temperatures
3	6.47 - 7.02 (upper level water vapour)	4 km	estimating regions of mid-level moisture content and advection; tracking mid-level atmospheric motion
4	10.2 - 11.2 (longwave IR)	4 km	identifying cloud-drift winds, severe storms, and heavy rainfall
5	11.5 - 12.5 (IR window sensitive to water vapour)	4 km	identification of low-level moisture; determination of sea surface temperature; detection of airborne dust and volcanic ash

The 19 channel **sounder** measures emitted radiation in 18 thermal infrared bands and reflected radiation in one visible band. These data have a spatial resolution of 8 km and 13-bit radiometric resolution. Sounder data are used for surface and cloud-top temperatures, multilevel moisture profiling in the atmosphere, and ozone distribution analysis.

NOAA AVHRR

NOAA is also responsible for another series of satellites which are useful for meteorological,as well as other, applications. These satellites, in **sun-synchronous, near-polar orbits** (830-870 km above the Earth), are part of the Advanced TIROS series (originally dating back to1960) and provide complementary information to the geostationary meteorological satellites (such as GOES). Two satellites, each providing global coverage, work together to ensure that data for any region of the Earth is no more than six hours old. One satellite crosses the equator in the early morning from north-to-south while the other crosses in the afternoon. The primary sensor on board the NOAA satellites, used for both meteorology and small-scale Earth observation and reconnaissance, is the **Advanced Very High Resolution Radiometer (AVHRR)**. The

AVHRR sensor detects radiation in the visible, near and mid infrared, and thermal infrared portions of the electromagnetic spectrum, over a swath width of 3000 km. The accompanying table, outlines the AVHRR bands, their wavelengths and spatial resolution (at swath nadir), and general applications of each.

NOAA AVHRR Bands

Band	Wavelength Range (µm)	Spatial Resolution	Application
1	0.58 - 0.68 (red)	1.1 km	cloud, snow, and ice monitoring
2	0.725 - 1.1 (near IR)	1.1 km	water, vegetation, and agriculture surveys
3	3.55 -3.93 (mid IR)	1.1 km	sea surface temperature, volcanoes, and forest fire activity
4	10.3 - 11.3 (thermal IR)	1.1 km	sea surface temperature, soil moisture
5	11.5 - 12.5 (thermal IR)	1.1 km	sea surface temperature, soil moisture

AVHRR data can be acquired and formatted in four operational modes, differing in resolution and method of transmission. Data can be transmitted directly to the ground and viewed as data are collected, or recorded on board the satellite for later transmission and processing. The accompanying table describes the various data formats and their characteristics.

AVHRR Data Formats

Format	Spatial Resolution	Transmission and Processing
APT (Automatic Picture Transmission)	4 km	low-resolution direct transmission and display
HRPT (High Resolution Picture Transmission)	1.1 km	full-resolution direct transmission and display
GAC (Global Area Coverage)	4 km	low-resolution coverage from recorded data
LAC (Local Area Coverage)	1.1 km	selected full-resolution local area data from recorded data